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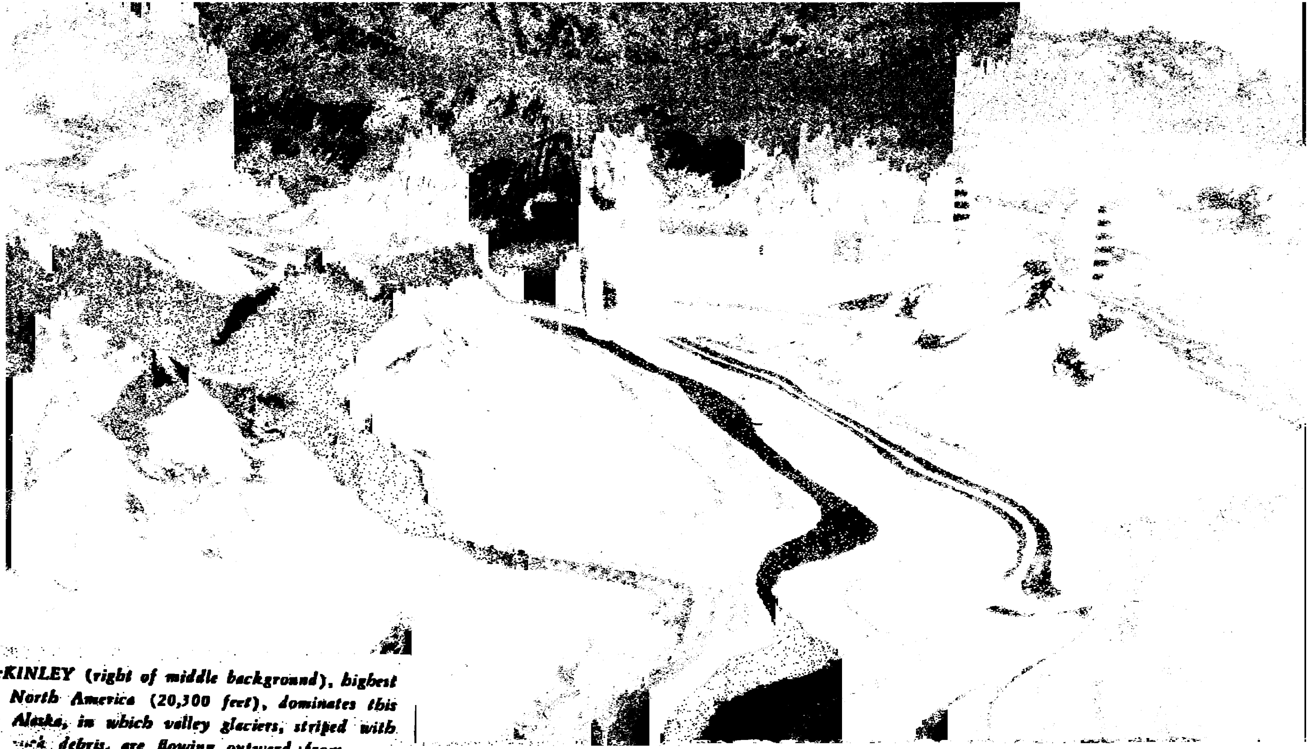

McKINLEY (right of middle background), highest in North America (20,300 feet), dominates this in Alaska, in which valley glaciers, striped with rock debris, are flowing outward from

Physical Geology

By

Chester R. Longwell
Henry Barnard Davis Professor of Geology
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THIRD EDITION

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Richard F. Flint

under the title

A Textbook of Geology
Part I—Physical Geology

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Second Printing, October, 1948

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Preface to the Third Edition

Both the first edition of *Physical Geology* (1932) and the second edition (1939) had a gratifying reception in numerous colleges and universities. Through continued use of the book in our own teaching, and through suggestions kindly offered by other users, we came to recognize that many parts of the treatment were in need of improvement. The present revision has not involved radical changes in major organization of the subject matter. A new unit, Chapter 2, has been added, the entire text has been critically analyzed, considerable sections of chapters have been reorganized and recast, and significant new data and concepts have been incorporated.

Nearly all instructors adapt a textbook to their own ideas of the most desirable order, emphasis, and method of presentation. This is altogether commendable; textbooks should be accepted not as rigid codes, but rather as aids in teaching, to be used with any desired flexibility. Whatever order of subjects is adopted, however, the authors are convinced that beginners in geology require at the start considerable orientation, to acquaint them with the major goals, the methods, and the basic outlines of a vast and unfamiliar field of knowledge. Elementary students can not be expected to acquire at once any adequate comprehension of geologic time; nevertheless this essential concept should be brought to their attention early and emphasized repeatedly throughout the course. The evolution of hypotheses into natural laws should be explained, with particular application to geologic problems now under attack. Geologic processes dependent on energy derived from the Sun should be differentiated from those reflecting internal forces, and the interaction between the two sets of processes should be pointed out. Unless the student is given a satisfactory preview of these and other major concepts, he is in danger of losing perspective in a maze of details.

These considerations have prompted an amplification of the introductory, orienting material in the text, by addition of the new Chapter 2. Transfer to this chapter of topics related to crustal movements serves a dual purpose: (1) the concept of isostatic balance in its relation to deformation of the Earth's crust provides an excellent illustration of the scientific method as applied to a geologic problem; (2) early introduction of this concept will give the student better comprehension of the references to uplifts and depressions of the land that are unavoidable in discussions of land sculpture, sedimentation, and other gradational processes. Once the significance of isostasy is grasped, it can be used as a logical working principle throughout the course. The interrelation of external and internal forces is so intimate that there can be no intelligent discussion of one set of forces and its effects without reference to the other set.

As in the earlier editions, definitions and elementary descriptions of all minerals and rocks considered in the text are brought together in appendices. This segregation does not belittle the importance of highly essential subject matter. On the contrary, the device serves a useful purpose that is twofold: it removes a considerable amount of bare definition and description from the body of the book, which can then be devoted more effectively to discussion of geologic processes and problems; and it concentrates, in the form of a short manual, the elementary explanations required for laboratory study. The arrangement can be adapted to varied plans that are used in beginning courses. It is particularly suited to the general program of a course that is suggested by the sequence of topics in the present edition of the textbook. This program, in brief outline, is as follows:

(1) Several exercises, including as much discussion as possible, devoted to a broad outline of physical geology, its viewpoints and methods, its objectives—both practical and philosophic, some major conclusions already attained, some outstanding problems now under attack.

(2) Study of common minerals and rocks, chiefly in the laboratory, using all available time until students have sufficient familiarity with the materials of the Earth's crust to comprehend the effects of geologic processes.

(3) Analysis of the gradational processes—erosion and sedimentation.

(4) Analysis of effects of internal forces—igneous activity, deformation, metamorphism.

(5) Explanation of land forms as joint products of internal and external forces.

(6) Mineral deposits.

As much field and laboratory study as possible or desirable will form part of the program.

In the present revision of the book, improvement of the illustrations has received particular attention. Most of the diagrams are either revised or wholly new, and many of the halftones are from new copy. Photographs in which small details have critical value are reproduced at "bleed" size.

Colored photographs have large value in geologic teaching, since tones of color bring out critical features and give a sense of reality that is largely lost in black-and-white views. It is not practicable to print numerous colored views in a textbook because of the large cost. However, the publishers of *Physical Geology* have provided a large set of Kodachrome slides showing geologic features that illustrate nearly all aspects of the subject. This set, accompanied by a manual which keys the slides directly to particular topics in the present edition, should be a valuable supplement in classroom presentation.

Direct responsibility of each author in preparing the third edition of the book is indicated as follows:

Chapters 1, 2, 3, 4, 10, 15, 16, 18, 19, and Appendix D by Longwell.
Chapters 5, 6, 7, 8, 9, 11, 20, and Appendix C by Flint.
Chapters 12, 13, 14, 17, 21, and Appendices A and B by Knopf.

Although acknowledgment is made with each figure, we wish to emphasize our gratitude to individuals and organizations for the use of illustrative material. In thanking numerous friends who gave help in improving the text, we are especially grateful to our colleague, Dr. John Rodgers, whose constructive suggestions applied to nearly all chapters.

C. R. L.
A. K.
R. F. F.

NEW HAVEN, CONN.,
June, 1948

Preface to the Second Edition

In the seven years since the first edition of *Physical Geology* was published the authors, through use of the book in their own teaching and from discussion with other teachers, have come to realize that some parts of the treatment should be recast and strengthened. Furthermore, during these years a large volume of fresh geologic literature has appeared, and distinct advances in several aspects of geology have been made.

The general plan of the book remains unchanged. Two new chapters have been added: Chapter 3, which gives particular attention to highly important movements such as creep and landsliding, all of which are included in the suggested term *mass-wasting;* and Chapter 5, in which the sculpturing of the lands by running water is treated as a unit, following a discussion of the principles of stream action in the preceding chapter. All the chapters have been thoroughly revised, and several have been completely reorganized.

Numerous cross-references are employed, with the object of coordinating the subject matter and eliminating needless repetition. The practical aspects of geology, such as its relation to soil conservation and industrial developments, are given particular attention. Technical terms have been eliminated as far as seems consistent with effective and clear treatment. Many new reading references are listed, and new field results are cited to illustrate principles. A special effort has been made to improve the illustrations: some of the figures used in *Outlines of Physical Geology* are included, as well as numerous photographs and diagrams that are new.

Because of the two new chapters, the numbering of chapters is considerably different from that in the first edition. Direct responsibility of each author is indicated as follows:

Chapter 1 by Knopf and Longwell.
Chapters 2, 3, 9, 14, 15, 17, 18, and Appendix D by Longwell.
Chapters 4, 5, 6, 7, 8, 10, 19, and Appendix C by Flint.
Chapters 11, 12, 13, 16, 20, and Appendices A and B by Knopf.

Although acknowledgment is made with each figure, the authors wish to emphasize their gratitude to individuals and organizations for the use of illustrative material.

C. R. L.
A. K.
R. F. F.

NEW HAVEN, CONN.,
January, 1939

Preface to the First Edition

Three years ago the writers helped revise Part I of the ***Textbook of Geology*** by Pirsson and Schuchert. Extensive changes were made throughout, and several chapters were entirely rewritten; but as the revised book remained under the authorship of the late Professor Pirsson, the revisers felt an obligation to retain as nearly as possible the original method of presentation. The present volume, which embodies not only ideas of organization and presentation developed by the work of revision and by constant use of the revised text, but also much new material, is the avowed successor to Pirsson's book. Probably the most valuable inheritance is the balance between subjects for which the original book has been widely commended. The general order of the final revision is retained, and in a few chapters some parts of Pirsson's scheme of presentation have been used, but only where they could be effectively adapted to the general plan of the present authors.

The organization of the subject matter assumes that students who use the book begin their study by making an acquaintance with the common minerals and rocks. This approach to the study of geology appears to be fundamental, since the operation of geologic processes, for example weathering, can not be appreciated without some knowledge of the materials involved. In their elementary course the authors devote a considerable time to a practical study of minerals and rocks before taking up any of the processes. For such a preliminary study the student needs adequate descriptions arranged in convenient form. To meet this need the present volume offers in two appendices the definitions and elementary descriptions of all the minerals and rocks considered in the text. Another appendix gives an introduction to the study of topographic maps. Such a special arrangement does not in any way obscure the importance of highly essential subject matter. On the contrary, the device serves a twofold useful purpose: it brings together, in the form of a short manual, the elementary explanations needed for laboratory study; and it removes a considerable amount of bare definition and description from the body of the text, which can then be devoted more effectively to the discussion of geologic processes and problems.

Preparation of the introductory chapter has received particular attention, since physical geology usually is the introduction to geology as a whole and therefore the student needs an orienting statement to acquaint him with the scope and the general viewpoint of the subject. The historical aspect of geology needs emphasis, and this purpose is served by sketching briefly the astronomical setting of the Earth and by implanting at once the concept of the immense length of geologic time. A large advantage is gained if the student can be made to appreciate at the outset that geologic features are studied not merely for their own sake but more particularly for the history they reveal.

With this keynote in mind, the attempt is made throughout the book to keep to matters that have a direct bearing on geologic history. The authors consider it a mistake to use the limited space in an elementary text for the exposition of scientific matters not directly a part of geology, however interesting they may be in themselves. For this reason the discussion of weather, of which at best only a smattering can be given in a geologic text, has been abandoned. Similarly a detailed discussion of the mechanics of glacier motion, which has inherent interest but is not essential to an understanding of land sculpture by glaciers, is dispensed with.

In brief summary, some of the outstanding features of the new book are the following: Weathering is treated as a unit subject, and its relation to general erosion is made clear.

The chapter is not entitled "Work of the Atmosphere," because all gradational processes, including stream erosion and glaciation, depend either directly or indirectly on the atmosphere. The importance of climate in determining soil types, as demonstrated by recent research, is emphasized. Wind action is discussed in a separate chapter, since the common practice of combining the subject with weathering leads to confusion. In the discussion of stream work the explanation of the erosion cycle under different climatic conditions is given more emphasis than in the final edition of the Pirsson textbook. The discussions of glaciation, igneous geology, and metamorphism are revised to take account of the most recent data. Modern researches in seismology, and particularly the late work on near earthquakes, are considered in the chapter on the Earth's interior. The relation of geologic processes to landscapes has been given special attention in nearly every chapter. As in the final edition of the Pirsson textbook, the treatment of the more complicated aspects of land forms is deferred until deformation and other processes have been explained in detail; but the old treatment has been entirely reorganized and much new material has been added. The final chapter gives a discussion of coal and petroleum, in addition to a short treatment of ore deposits, and it emphasizes that our understanding of these mineral resources depends on geologic principles developed earlier in the book.

Discussion of many major subjects in geology leads quickly into the fields of inference and theory. Some elementary textbooks, probably for the purpose of simplifying and shortening the treatment of debatable matters, make dogmatic statements not warranted by existing knowledge. In the opinion of the present writers this practice does a disservice to the science. The teacher loses nothing by stating frankly what is not known; on the contrary he can in this way gain the student's confidence and stimulate his interest. The cause of rock exfoliation has not yet been proved; we can not give any quantitative evaluation of wind erosion; the origin of igneous rocks and the causes of crustal movements are great mysteries. Since almost every aspect of geology teems with fascinating problems, why should an elementary text give the false impression that most of the Earth's secrets are known? It is the aim of the present volume not only to state the essential facts of physical geology, but also to indicate the limits of knowledge and to direct the student's attention to the vast field awaiting exploration.

Most of the illustrations used are new. Numerous block diagrams, specially drawn for this book, are employed to explain the genesis of structural features and land forms. None of the figures are used simply for embellishment; all have been chosen for their geologic significance. Duplication is avoided by referring both forward and backward to figures that illustrate more than one subject or principle. Special care has been used in preparing the legends of the illustrations, with the purpose of centering attention on significant features and making the figures intelligible without direct reference to the text.

With the idea of welding the chapters together and showing the interrelation of various subjects, cross-references are used freely throughout the book.

Direct responsibility of each author in preparing the text is indicated in the following division by chapters:

Chapter 1 by Knopf and Longwell.
Chapters 2, 7, 12, 13, 15, 16, and Appendix D by Longwell.
Chapters 3, 4, 5, 6, 8, 17, and Appendix C by Flint.
Chapters 9, 10, 11, 14, 18, and Appendices A and B by Knopf.

The authors are under obligation to many individuals and organizations for the use of photographs. Proper acknowledgment is made with each halftone.

C. R. L.
A. K.
R. F. F.

New Haven, Conn.
January 1, 1932

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CHAPTER 1

GEOLOGY, THE SCIENCE OF THE EARTH

The science of geology strives for the full answer to questions about the Earth and its history. What are the materials that compose the Earth, not merely in its outer part but also in the deep interior? What is the meaning of mountains and other landscape features? How and when did the Earth itself originate, and what has been the full story of life in its many forms? The philosophers of the ancient world wrestled with these problems, although they contributed little of lasting value to the science; geology had its first consistent development in modern times, with the advance of science generally. The practical problems of mining stimulated the development of modern geology, but the growth of the science to its present substantial proportions was largely the result of the intellectual curiosity of the human mind. The principles thus discovered have been applied to practical ends, and geology has repaid a thousandfold its early debt to mining.

Geology enlists all the other natural sciences—physics, chemistry, astronomy, and biology—in examining the various aspects of the Earth and its history. A fitting introduction to the study considers the evidence from astronomy as to the place of the Earth in the universe.

THE EARTH AS A MEMBER OF THE SOLAR SYSTEM

The Earth is one of a group of nine known planets that revolve around a common central orb, the Sun (Fig. 1). Some of them, like Jupiter, are much larger than the Earth; some, like Mercury, are smaller. Two are nearer the Sun, the others are farther away; the distances range from about 25 million miles for Mercury to thousands of millions of miles for the three outermost planets. They all revolve around the Sun in nearly a common plane, which is slightly inclined to the Sun's equator.

The mass of the Sun is 332,000 times that of the Earth. All the planets and their satellites combined make up only a small fraction of 1 per cent, in mass, of the Solar System. The Sun contains considerably more than 99 per cent of the total mass, and therefore it is the overwhelmingly dominant member of the system. Carrying with it the planets and their moons, the comets, and the asteroids, the Sun is traveling through space at the rate of 12 miles a second, almost directly toward the bright star Vega.

The Earth, therefore, is a very insignificant fraction of the Solar System, and for that matter the Solar System itself is but an insignificant fraction of the universe. The Sun is a modest star in a stellar system whose members are numbered in thousands of millions. At immense distances beyond this system—"our galaxy" as it is currently called in astronomy—are other galaxies, perhaps a hundred million of them, which are gigantic systems comparable to ours in size and in number of stars. Some of these systems are so remote that the light

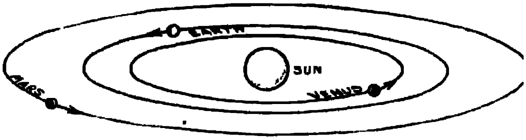

FIG. 1. Planets revolve about the Sun almost in the same plane, and all in the same direction, as suggested by the orbits of three of them. The orbits are nearly circular, but in the figure they appear strongly elliptical because the observer is looking down upon them obliquely. The diagram is not drawn to scale.

from them, traveling 186,000 miles a second, takes hundreds of millions of years to reach us. Nevertheless, throughout this vast extent of space the same general physical laws that we know on the Earth hold sway. Gravitation operates in the same manner; light is transmitted everywhere by vibrations of the same kind; the spectroscope tells us that the same chemical elements occur in the Sun and the distant stars as on our Earth. Moreover, the meteorites that our planet gathers in its journey through space are made of substances identical with those found on the Earth. The processes at work in the Sun and stars, however, far exceed in intensity those active on and in the Earth, because of the enormously greater temperatures and pressures that Nature has at her command in the stars; but they do not differ in kind. Consequently there is a unity of law and a uniformity of material throughout space, and we feel justified in assuming that facts and reasoning derived by astronomical study of the other heavenly bodies can also be applied in the study of the Earth. One of the most impressive demonstrations of this principle was afforded by the discovery of helium: first recognized as a constituent of the Sun, 93 million miles away, this element was found nearly 30 years later on our planet.

The Earth and the other planets were born of the Sun, as has been recognized since the latter half of the eighteenth century when the nebular hypotheses of Kant and Laplace were formulated. How the incandescent matter from the Sun came to form the planets is a problem on which ideas are still far apart. The birth time of the Earth was at least 2000 million years ago, and since then the Earth has continued to have an eventful history, which it is the special task of geology to decipher. The record of a vast span of time is written in the rocks which make up the visible portion of the planet's outer shell. Just as in human history, so in the Earth's history, the earlier records are scarce and nearly illegible, but the later records are much more abundant and more easily read. That portion of the Earth's history which is written in the rocks is referred to as *geologic time*. For this portion of the history we have positive information, and no matter how fascinating speculations concerning the Earth's origin may appear, the foundations of geology do not rest on them; on the contrary, the most probable fact concerning the origin of the Earth, that it was born at least 2000 million years ago, rests on geologic evidence supported by the study of radioactive minerals found in the Earth's crust.

The geologic record shows that the Sun has been supplying light and heat to the Earth at a nearly uniform rate during hundreds of millions of years. Prior to the development of nuclear physics in the present century, the sciences could not adequately explain how the Sun has been able to maintain this prodigal expenditure of energy. It is now thought that its fires are stoked through atomic energy in the building up of heavier from lighter chemical elements. According to one reputable theory, much of the heat is released during the development of helium from hydrogen, with carbon playing an essential part as catalyst.

MAJOR DIVISIONS AND FEATURES OF THE EARTH

Form of the Earth. Quite aside from its surficial irregularities, the Earth is not a true sphere; it is a spheroid flattened at the poles, so that the axis on which it rotates is shorter by about 27 miles than the equatorial diameter. This equatorial bulging of the Earth, as well as that of the other planets, is fully explained by the principles of celestial mechanics. It depends on the centrifugal force due to rotation and on the internal constitution of the planet (Chap. 18).

Atmosphere and Hydrosphere. The solid body of the Earth has two envelopes which are of the utmost importance in geology. One of these, the *atmosphere*, is made up of gases; it extends in appreciable quantity more than 100 miles above the surface of the lands, but in the outer part the gases are extremely rare. Although the material in the atmosphere makes a very small percentage of the entire Earth, it is one of the most important geologic agents because of its chemical and physical activity. Oxygen and carbon dioxide are active ingredients that play extremely important roles in geology. The more abundant nitrogen has little chemical activity.

The other envelope, the *hydrosphere*, is not continuous, although the sea covers nearly three-fourths of the Earth's surface, and water is widespread on the lands also, either in lakes and streams or filling openings in the soil and in bedrock. So vast is the volume of water that, if all surface inequalities of the globe were leveled off, the sea would be universal, with a depth of more than a mile and a half. Water, with the aid of the atmosphere, has been the most powerful agent in causing changes on the surface of the lands throughout geologic time.

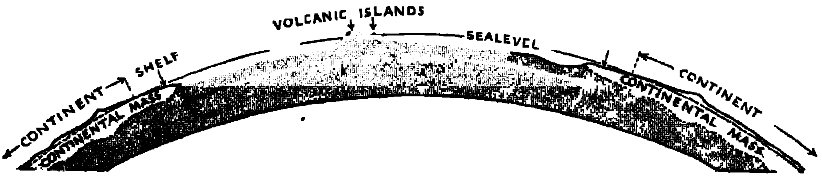

FIG. 2. Segment of the Earth, showing major features of relief The vertical scale is greatly exaggerated.

In general, the light gases of the atmosphere overlie the denser water of the hydrosphere, which in turn lies above the much denser material of the solid Earth. Owing to the irregular surface of the globe, large areas project above the hydrosphere and are directly accessible to observation.

Continental Masses and Deep-Sea Basins. Of the 197 million square miles making up the surface of the globe, 71 per cent is covered by the interconnecting bodies of marine water; the Pacific Ocean alone covers half the Earth and averages near 14,000 feet in depth. The *continents*—Eurasia, Africa, North America, South America, Australia, and Antarctica—are the portions of the *continental masses* rising above sealevel. The submerged borders of the continental masses are the *continental shelves*, beyond which lie the deep-sea basins (Fig. 2).

The oceans attain their greatest depths not in their central parts, but in certain elongated furrows, or long narrow troughs, called *deeps* (Chap. 11). These profound troughs have a peripheral arrangement (Fig. 2), notably around the borders of the Pacific and Indian oceans.

The position of the deeps near the continental masses suggests that the deeps, like the highest mountains, are of recent origin, since otherwise they would have been filled with waste from the lands. This suggestion is strengthened by the fact that the deeps are frequently the sites of world-shaking earthquakes (Chap. 16). For example, the "tidal wave" that in April, 1946, caused widespread destruction along Pacific coasts resulted from a strong earthquake on the floor of the Aleutian Deep.

The topography[1] of the ocean floors is none too well known, since in great areas the available soundings are hundreds or even thousands of miles apart. However, the floor of the Atlantic is becoming fairly well known as a result of special surveys since 1920. A broad, well-defined ridge—the Mid-Atlantic ridge—runs north and south between Africa and the two Americas, and numerous other major irregularities diversify the Atlantic floor. Closely spaced soundings show that many parts of the oceanic floors are as rugged as mountainous regions of the continents. Use of the recently perfected method of echo sounding (Chap. 11) is rapidly enlarging our knowledge of submarine topography. During World War II great strides were made in mapping submarine surfaces, particularly in many parts of the vast Pacific basin.

The continents stand on the average 2870 feet—slightly more than half a mile—above sealevel. North America averages 2360 feet; Europe averages only 1150 feet; and Asia, the highest of the larger continental subdivisions, averages 3200 feet. The highest point on the globe, Mount Everest in the Himalayas, is 29,000 feet above the sea; and as the greatest known depth in the sea is over 35,000 feet, the maximum *relief* (that is, the difference in altitude between lowest and highest points) exceeds 64,000 feet, or more than 12 miles. The continental masses and the deep-sea basins are relief features of the first order; the deeps, ridges, and volcanic cones that diversify the sea floor, as well as the plains, plateaus, and mountains of the continents, are relief features of the second order. The lands are unendingly subject to a complex of activities summarized in the term *erosion* (p 29), which first sculptures them in great detail and then tends to reduce them ultimately to sealevel. The modeling of the landscape by weather, running water, and other agents is apparent to the keenly observant eye and causes thinking people to speculate on what must be the final result of the ceaseless wearing down of the lands. Long before there was a science of geology, Shakespeare wrote "the revolution of the times makes mountains level."

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[1] *Topography* relates to the surface forms of an area, such as heights, depressions, and slopes. See Appendix C.

The origin of the continental masses and the deep-sea basins is still an unsolved problem. Have they existed in their present form and position since the beginning of geologic time? Have deep-sea floor and land changed places? If the continents have been permanent features of the globe for some 2000 million years, or even for a much smaller span of time, why have they not been worn down to or below sea-level? Manifestly there is some restorative force acting from within the Earth that counteracts the destructive effects of erosion acting from without the Earth. To these questions we shall return in later chapters.

ROCKS, THE PRIMARY DOCUMENTS OF GEOLOGY

The outer zone of the Earth—the zone on which we live and the one directly accessible to our observation—is generally known as the *crust*. There is some objection to this term, because it is inherited from an obsolete hypothesis, which erroneously assumed that all the Earth except a thin outer shell is molten liquid. However, "crust" is a convenient and popular term, and we shall use it with the understanding that it does not suggest the condition and origin of the Earth's interior (Chap. 18).

Any understanding of the Earth must begin with some knowledge of the substances that compose it.[1] We can of course examine directly only a thin rind of the globe, and therefore all information about the vast interior portion must depend on indirect evidence. However, our most active interest lies in the part that can be explored directly; and this part in itself provides an almost limitless field of study. It is a zone composed of rocks and their constituent minerals. Even to the casual observer it is evident that these materials, exposed in cliffs or in road cuts, tunnels, and other artificial excavations, are highly varied in character; and accordingly it may appear that only a specialist can hope to gain any adequate acquaintance with rocks. Fortunately, rocks of relatively few types make up the greater part of the Earth that is visible, and therefore any educated person can learn without great difficulty to recognize most of the rock masses he may see in the Alps, the Rocky Mountains, or elsewhere in his travels. A study of variations in rock types or of the more detailed features in minerals must of course be left to men specially trained for the task.

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[1] Some familiarity with the common minerals and rocks, to be acquired from the descriptions in Appendices A and B (pp. 546, 560) and from laboratory study of typical specimens, is essential for understanding the processes discussed in the following chapters.

HAWAIIAN VOLCANO OBSERVATORY.

Fig. 3 Igneous rock in the making The dark-colored rock is solidified lava; the white band is a stream of fluid lava, flowing toward the observer The Alika flow, Mauna Loa volcano, 1919.

The practical value of knowing the various kinds of rocks is readily apparent from the viewpoint of the professional geologist or the mining engineer. In the search for ores and for petroleum, in the selection of sites for great dams to make storage reservoirs, or in constructing an intricate subway system for a large city, knowledge of rocks and their peculiarities is a vital necessity. But there is a much broader interest in the subject—an interest felt by every intelligent traveler. A visitor to the slopes of Vesuvius or of Mauna Loa sees masses of dark, slaggy rock. It is fairly obvious, even if fluid lava is not actually seen, that this dark material was once liquid and flowed down the slopes as a red-hot stream (Fig. 3). Continued investigation would convince the traveler that the "fire-made" or *igneous* rocks are common in many lands and have been formed at many different dates; they constitute an important part of the bedrock beneath us In fact, *granite,* which is overwhelmingly the most abundant type of rock in the foundations of the continents, was made, during various periods of geologic time, by slow cooling and solidification of liquid rock material deep beneath the Earth's surface. Therefore the igneous rocks, recognized as one of the main classes of rocks in the Earth's crust, are of two general kinds: (1) those formed on the Earth's surface, from molten material forced up from the depths; and (2), those formed beneath the surface, some of which have been exposed to view by long-continued erosion (Fig. 198, p. 290).

Again, a brief examination in parts of the Andes, the Alps, or other high mountains reveals layers of compacted sand and mud that include an abundance of sea shells. These ancient shell-bearing rocks have modern equivalents in the muds and sands of river deltas and on sea floors. Wind, water, and other agents are constantly moving and spreading out rock fragments or *sediments*, which in time become compacted and cemented to form the *sedimentary* rocks (Fig. 11; Chap. 12). Three-fourths of the land area of the Earth is directly underlain by rocks of this kind. In most places, however, these rocks form a comparatively thin cover—a few hundreds or thousands of feet thick—lying on rocks of other kinds, predominantly granite. The sedimentary rocks reveal important events in the Earth's history, because they contain remains of plants and animals and also many other features that give a record of conditions existing when the rocks were formed.

In the cores of old mountain ranges, and also around the borders of igneous masses, there are rocks of another class known as the *metamorphic*. These rocks, originally either igneous or sedimentary, have had their original characters greatly changed by temperature, pressure, and other factors acting within the Earth's crust. Metamorphic rocks commonly are strongly banded (Fig. 282, p. 430), and some, such as slate, can be readily split into thin sheets or flakes. Rocks subjected to very high temperature or to extreme pressure have had their primary characters entirely obliterated, in which case it may be impossible to tell whether the rocks were originally igneous or sedimentary.

In summary, all the rocks in the Earth's crust are grouped, according to the way they were formed, into three main classes: igneous, sedimentary, and metamorphic rocks.

THE VIEWPOINT OF GEOLOGY

The grouping of rock masses according to the mode and place of their origin is the first step in the fascinating task of unraveling the ancient history of a region. In all its many aspects, the study or practice of geology recognizes this fundamental interest in past events. But just as the proper understanding of human history requires some knowledge of present-day social, economic, and political conditions, so the deciphering of events in the geologic past demands an acquaintance with processes still operating on and within the Earth. Rocks are not inert monuments to conditions and forces that no longer exist. At active volcanoes we may observe all stages in the development of new rocks from molten material (Fig. 3). In the muds of river deltas and on sea floors we find the modern equivalents of ancient beds, now greatly distorted and eroded, which furnish the shells and other fossils so common in high mountains and plateaus. Rivers, glaciers, and other surface agencies are etching and slowly wearing away the lands. In some continents the land is being lifted up at a measurable rate, and there is evidence that some mountains are actively growing. Thus we are actual witnesses to a constant struggle between titans: some that work at the surface, striving to tear down the rocky continents, and others within the Earth that persistently oppose the leveling process. Rocks are being destroyed and others are forming to replace them. Activities that can be seen and analyzed have been persistent throughout geologic time, and accordingly we may use the present as a key to the past. It is the conclusion of modern geology that the great mountain ranges, deep canyons, and other major features of the Earth came into existence not through great *catastrophes*, but as the cumulative results of ordinary processes still active, operating slowly through enormously long intervals of time For example, some people who are not acquainted with geology suppose that the Grand Canyon of the Colorado River, more than a mile in depth and 200 miles long, is a great crack in the Earth, formed abruptly at the time of a violent earthquake. Geologists are convinced that the canyon was fashioned slowly, during a long interval of time, through the cumulative effects of erosion by running water, which is still in operation. This concept, that major features of the Earth have been developed through long-continued, nearly *uniform* action of forces now operating, is known as the *uniformitarian* principle. It is less spectacular than views of the *catastrophists*, but it is firmly established by a great body of geologic evidence.

The broad, philosophical aspect of geology rests on a secure foundation only because of patient effort by generations of workers in all lands. A vast array of field evidence has been accumulated. The great mountain ranges have been fruitful fields for geologic investigation, since they furnish the finest exposures of rock formations. But no sources of information are neglected. Geologic features everywhere are examined closely in the field and represented accurately on maps.

Mines give opportunity to explore beneath the surface, and deep wells drilled for water or oil yield valuable data. Samples collected from sea floors in many parts of the world reveal diverse kinds of deposits that are accumulating to form sedimentary rocks of the future. Modern instruments devised to record earthquake waves and to measure values of gravity have made it possible to draw important conclusions regarding the invisible interior. Slowly but surely the Earth is giving up many of its secrets to the inquisitiveness of man.

Of necessity, a field so broad as the study of the whole Earth calls for a division of labor among specialists. One branch of the study (*stra-tig'ra-phy*) is devoted chiefly to the rock formations laid down in former seas, lakes, and streams; another (*pe-trol'o-gy*) to the character and origin of rocks of all kinds; a third (*economic geology*) to the mineral veins and other deposits of economic value. Other special branches of geology are concerned with the deformation of rocks by folding and fracturing (*structural geology*), the various land forms sculptured by surface agencies (*geomorphology*), the fossils entombed in rocks (*paleontology*), and the minerals that make up rocks of all kinds (*mineralogy*). A general introductory study can attempt only a survey of methods and conclusions, without giving an intimate acquaintance with any of the special fields. However, such a general study of geology should equip the student to recognize and interpret many of the major geologic features that he sees about him. In particular, the varied landscapes take on a new meaning and interest when they are seen as clear records of past events. Land forms are products of a complicated interplay between crustal movements on the one hand and sculpturing agents on the other. For example, fashioning of the Matterhorn and similar neighboring peaks was made possible by great uplift of a rock mass to form the Alpine chain; the cold climate and steep slopes on this mass favored the development and movement of glaciers, which have carved the summit into its present rugged form (p. 190). The shape of every landscape feature depends also on the kind of rock from which it was made and the length of time since the sculpturing began. Therefore, although there is no fixed order in which geologic subject matter must be considered, full appreciation of Earth features requires some preliminary knowledge of the common rocks and of their behavior under the forces active at or near the Earth's surface.

The first part of geologic study, devoted to the materials and features of the Earth and the processes now in operation, is called *physical geology*. *Historical geology*, which reconstructs the story of past physical events and also the record of life on the Earth, uses as keys the principles developed in physical geology and also some of the principles of biology.

The Earth and its history are fascinating subjects in themselves, but the interest is largely academic unless it is related to man and his welfare. The Earth is our home; we obtain our physical nourishment from its soil; we are affected directly by its storms and floods, its volcanoes and earthquakes, the uplift and sinking of lands, the storing of metals, coal, and oil in the rocks. Moreover, as the history of the Earth continues to unfold, we see unmistakably that the events of long ages led up to the appearance of man.

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CHAPTER 2

THE METHOD AND SCOPE OF GEOLOGIC STUDY

The importance of learning facts about the Earth, as a fundamental basis of geologic study, has been emphasized (p. 6). Prior to the nineteenth century many natural philosophers announced sweeping conclusions on geologic matters based largely on assumptions, in ignorance of truths that could be learned by observation in the field. There resulted rival schools of thought, each devoted to some speculative concept. Gradually it came to be realized that speculations, however ingenious and popular they may be, acquire real value only as they successfully meet repeated tests of new information. In the early part of the nineteenth century many students of the Earth, impatient with the earlier vogue of accepting hypothetical explanations "on authority," set about the laborious task of examining and describing the vast array of geologic features on and beneath the Earth's surface. That was a veritable epoch of fact finding, an extremely important stage in the growth of geologic science However, science is not a mere collection of facts; it is *systematized knowledge*, and its goal is the discovery of natural laws, to which the facts bear witness. This point of view, which is essential for any real understanding of aims and accomplishments in science, merits further discussion.

THE SCIENTIFIC METHOD

The scientific attack on a problem follows an orderly procedure, in a succession of steps suggested by the following questions: (1) What are the known facts in the problem? (2) How do these facts appear to be related? (3) How may this relationship be explained? (4) Is this explanation supported by all obtainable new information? At the start there may be only scattered facts, dubiously related. Intelligent guesses, or *hypotheses*, are set up in attempting an explanation. It then becomes obvious that certain specific information, if it can be obtained, will either support or weaken a particular hypothesis. Thus the scientific guessing game stimulates further search and inquiry, resulting in additional factual knowledge, which may eliminate some of the hypotheses and point to one as the most promising, though perhaps with some modifications in its details. If, after a considerable body of new pertinent data has been built up, the favored hypothesis remains, in its essentials, as the most adequate explanation, this concept then begins to acquire the more reputable status of *theory*. As supporting evidence becomes more and more convincing, the concept gradually emerges from the realm of theory into established natural law.

Development of a vague hypothesis to the status of a major scientific principle is seldom if ever accomplished by one individual. Progress in such matters usually has been slow, requiring contributions of numerous workers through more than one generation. Newton had a major role in formulating the law of gravitation, but, contrary to popular belief, he did not originate the entire concept. Kepler, Galileo, and others had grasped parts of this fundamental principle nearly a century before Newton's time. Another illustration, in the field of biology, may be drawn from the "fact-finding epoch" in the late eighteenth and early nineteenth centuries. Voluminous descriptions of modern and fossilized animals and plants had focused attention on relationships and diversities among living forms, and more than one naturalist came to entertain the concept that species have originated by evolutionary development. In 1815 the French scientist Lamarck published his views on this subject, including his well-known suggestion that structural changes in animals have come about slowly through use of bodily parts in response to need, and that characters thus acquired by individuals are inherited and perpetuated in their offspring. According to this view the giraffe's neck became elongated by stretching, through many generations, in efforts to reach the foliage on which the animals depended for food. This hypothesis has not survived the tests of further study. Darwin later developed the rival concept of evolution by "natural selection" of types best fitted to survive under given conditions. This view, in considerably modified form, is still a part of biologic theory, although the exact mechanism by which evolutionary changes take place remains an unsolved problem. However, the evidence that such changes actually have occurred is now overwhelming, and organic evolution has long been accepted as a natural law.

In physics, chemistry, and biology, laboratory experiments are an important tool in the scientific approach to a problem. Experimentation is valuable as an aid in solving some kinds of geologic problems also. In general, however, geology advances through reasoning based on accurate observations. Every geologic feature—a volcanic mountain, a layer of limestone, or a deposit of coal—presents a problem of origin; when and by what means did it come into being? The answers are to be sought through studies of two kinds: (1) by examination of the feature itself, in a search for clues that may suggest the time and mode of origin; (2) by a close study of processes, still active on the Earth, which may have played a part in fashioning the feature as we now see it. Thus we seek to read the complete history of every geologic feature and ultimately of the Earth itself. Because of this major objective of the study, geology is often called the *historical science.* A satisfactory introduction to the subject does not stop with an outline of chapters in the history that can be deciphered with fair assurance. The conclusions are significant only if the methods by which they have been reached are fully understood. Like any other growing subject, geology extends far beyond the lighted zone of proved fact. It has also a large twilight zone of inference and probability, into which the full light of investigation continues to spread slowly; and beyond this a region of darkness, relieved only by flashes of speculation. Thus a fascinating field for future exploration awaits those who have the zeal and proper equipment for the venture.

THE CONTINENTS; MOVEMENTS OF THE CRUST

A brief discussion of one major problem that is still partly in the twilight zone will illustrate the scientific method as it is employed in geology and will at the same time give the background needed for an understanding of geologic processes described in later chapters. This problem concerns the relation between continental masses and deep-sea floors, mentioned briefly on page 6. Terrestrial life, including mankind, depends on the continued existence of large areas above sea-level. Do the present continents represent merely accidental irregularities of the Earth's surface, doomed to eventual destruction by the forces of erosion? Is there constant danger that one or another continent may disappear permanently by subsiding several thousands of feet while part of the deep-sea floor rises in equal measure? Does geologic evidence indicate that lands were once deep-sea floors?

Equilibrium in the Crust. Examination of the rocks that make up the Earth's crust has brought out a highly significant fact. Under the continents, rocks that approximate granite in their composition are predominant. In islands that rise from the floors of the oceans, however, the rocks characteristically have the composition of dark basalt. Since thousands of islands that dot the vast Pacific area are the summits of volcanic mountains, their composition seems to represent that of the Pacific floor as a whole; if this floor were in any large part made of granite, some of the materials brought above sealevel by volcanic action would have granitic composition, just as granitic material appears prominently in many great volcanic mountains on continents.

Volume for volume, granitic rock is about 10 per cent lighter in weight than basaltic rock. Therefore an obvious hypothesis is suggested: great blocks in the crust made of low-density rocks, if they are to rest in balance with adjacent blocks that have higher density, must have larger volume and hence will rise to greater height. Implications of this hypothesis are so important that a more satisfactory checking of basic facts is demanded. Scattered oceanic islands represent only a tiny percentage of sea floors; is their testimony adequate? Fortunately, modern geophysical study of earthquakes yields accurate information on the nature of the rock traversed by the elastic impulses that pass from the point of origin to recording stations (Chap. 16). Such impulses pass through ocean floors with the high velocities characteristic of elastic waves transmitted through basaltic rocks. On the other hand, beneath continents the impulses from earthquake shocks move with the lower velocities corresponding to those determined for granitic rocks. Thus the geophysical evidence convincingly supports direct observations on the compositions of continental masses and sea floors. Moreover, the geophysical studies suggest strongly that the light-weight rocks under continents form a plate not more than a few tens of miles thick, underlain by heavy rock similar to that under the oceans (Fig. 4). This conclusion is strengthened by the presence on the continents of great fields of dark basaltic lavas, with strikingly uniform composition. These lavas, which have welled up through great

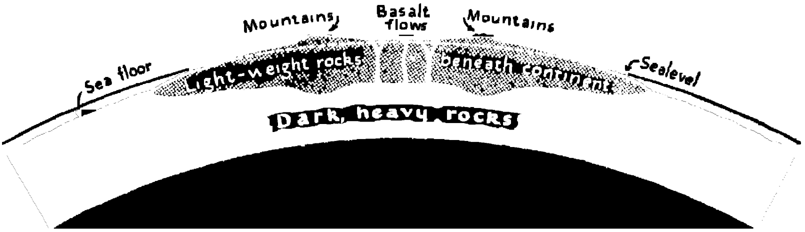

FIG. 4. Concept of a continent as the upper surface of a plate made of light-weight rocks, "floating" high above the heavier rocks on which it rests. Mountains are above the thickest parts of the plate, a plateau above a somewhat thinner part, low plains and continental shelves above still thinner parts. Great eruptions of basaltic lavas are represented as rising through the entire thickness of the continental plate from the continuous substratum of heavy rock (p. 333). Vertical scale greatly exaggerated.

fissures, supposedly had their origin in fused portions of the basaltic "subcrust" which everywhere underlies the granitic plate forming a continental mass (Fig. 4).

Thus it appears that the continents are not chance irregularities on the Earth; they are the upper surfaces of light-weight masses which float high in a denser medium, much as an iceberg rises above the water in which it is immersed. The principle of equilibrium is further illustrated in Fig. 5. Blocks of copper, with specific gravity 8.9, will float in mercury, which has specific gravity 13.6. The heights to which blocks rise above the surface are proportional to the depths of immersion. This suggests at once that in great mountain belts, such as the Andes and the Alps, the continental plate of light-weight rock is considerably thicker than under areas of low plains; the thicker segments are buoyed to greater height, and thus the great mountain masses are not loads on the Earth's crust—they rest in equilibrium, as do the higher blocks in Fig. 5.

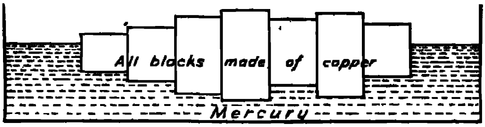

FIG. 5. Copper blocks, equal in cross section but unequal in length, float in mercury. They sink to unequal depth and also rise to unequal height. According to this conception of isostasy, a continent is the upper surface of a light-rock shell, essentially uniform in density but with variations in thickness. The thinner portions form low plains, whereas thicker parts project upward as plateaus and mountains. (See Fig. 4.)

The illustration of blocks floating in a liquid may appear highly artificial as applied to continental masses, since the Earth's crust is rigid. However, rock, which appears very strong in man-made structures, is weak and yielding in masses of continental dimensions. The Earth does not have the form of a cube or other angular figure, for a good mechanical reason. If the Earth could by some powerful force be greatly distorted and the force were then removed, physical law would quickly bring restoration to a figure of equilibrium. This figure would be essentially a sphere if there were no rapid rotation; but spinning on an axis makes the figure an oblate spheroid, with the degree of flattening determined by the rate of spinning. Similarly if the continental mass shown in Fig. 4 could be forced down until its surface was below sealevel, and then released, the familiar Archimedes' principle, or law of buoyancy, would cause slow deformation of the rocks beneath the continent until the surface arose to the proper height for equilibrium. The term *isostasy* (ī-sŏs′tȧ-sĭ)—from the Greek, meaning "equal standing"—is used for this condition of balance between segments of the Earth's crust.

Evidence in Glaciated Regions. It is not an idle speculation to suggest a force capable of bending a continental mass downward. More than once the lands that are now in the temperate zones have been chilled until great ice sheets thousands of feet thick covered millions of square miles (Chap. 9). These vast sheets of ice were made of water evaporated from the seas and precipitated as snow on the lands. Withdrawal of so much sea water lightened the load on all sea floors and concentrated this load on a much smaller total land area. After climates again grew warmer and the latest North American ice sheet vanished—only a few thousands of years ago—the sea invaded large areas in southeastern Canada and northeastern United States, but was excluded by later uplift of those areas. Emerged shorelines, associated with beach deposits that contain marine shells and bones of dolphins and whales, are found in the region of Lake Champlain and Montreal several hundred feet above present sealevel. Similar evidence is found in Scandinavia and Finland, which also were buried under thick ice during this glacial age (Fig. 225, p. 354). The inference is that the immense loads of glacier ice depressed the lands. After the loads were removed some time was required for restoration of balance by slow flowage in the rocks at depth. Before adjustment had progressed far, the sea came in to cover the depressed areas; and as balance was gradually restored by regional uplift the sea was expelled. Although it is thousands of years since the ice disappeared, the adjustment is not yet complete, as parts of Scandinavia are still rising at the rate of 2 or 3 feet in a century (p. 353) and the Great Lakes region in North America is undergoing measurable tilting by continuing uplift in Canada, where the ice load on this continent was greatest.

Isostatic Adjustment. The history of glaciated regions provides striking confirmation of isostasy in the Earth's crust. Shifting of large masses from one place to another, as in the transfer of sea water to form ice sheets on land, causes a change in the form of the Earth in order to restore equilibrium between loaded and unloaded areas. The growth and wasting of ice sheets has caused large but comparatively brief changes in distribution of the load of water on the Earth. Other geologic processes bring about equally large and more permanent transfers of mass. For example, great chains of mountains such as the Rockies are being eroded vigorously, and the resulting debris is being carried by streams to lowlands and sea floors hundreds or thousands of miles away. We see evidence that the present Appalachian ridges represent mere stumps of once-great chains from which erosion has removed vast quantities of rock. There is abundant evidence also that, as the agelong destruction of the Appalachian chains progressed, the region was repeatedly arched up (pp. 474, 500), just as the glaciated areas of continents were uplifted on removal of the load of ice.

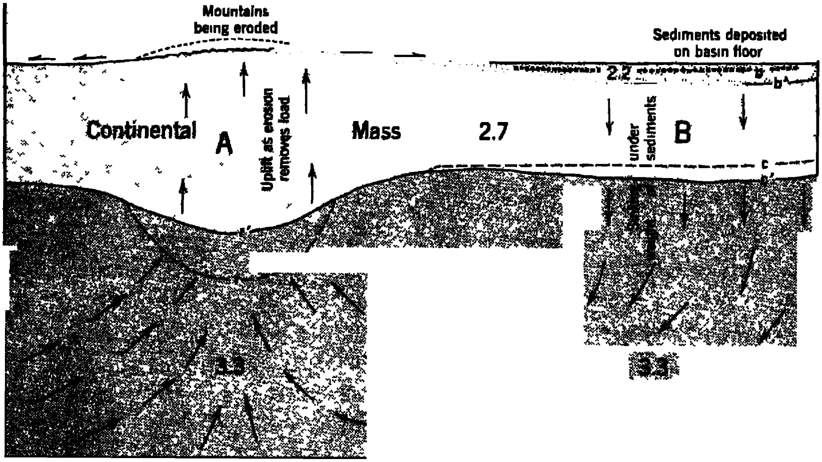

FIG. 6. The high mountain block *A* is eroded, the material removed is carried to low levels, and part of it is deposited on the low block *B*. When deposition began, the top of block *B* was at *b*, its base at *c*. Under the load of accumulating deposits (density the top of the block is depressed to *b′*, its base to *c′*. As load is removed from *A*, uplift of the block occurs, with the slow flowage of material at depth to maintain balance between the loaded and unloaded blocks. As the block rises, the space between and *c′*, formerly filled with rock of density 2.7, comes to be occupied by rock of density 3.3. If the average thickness of this space is 9000 feet, the average thickness of lighter rock eroded from the mountain area must be 9000 feet + 2000 feet (the average measure of lowering from the original surface), since the ratio 2.7:3.3 = 9:11.

Therefore the principle of isostatic balance in the Earth's crust is supported by impressive evidence and is winning acceptance as a natural law.[1] However, large questions remain, and some aspects of the problem are still in the realm of hypothesis. What is the mechanism of adjustment when great loads are shifted? If a highland con-

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[1] Studies of the lateral attraction exerted by great mountain masses, and of precise gravity values determined at numerous locations, widely distributed, give further strong support to the concept of isostatic balance in the Earth's crust. Space required for a satisfactory explanation of this evidence is not available here. See Reading Reference 2, page 27.

tinues to rise as material is removed from it by erosion, how can a mountain chain ever be reduced to low altitude? An attempt to answer these questions is represented in Fig. 6. It is well known that temperature in the outer part of the Earth rises with increasing depth. Some tens of miles down the rocks must be many hundreds of degrees hotter than at the surface. Although at such depths the high pressure tends to increase rigidity, it is supposed that the high temperature must make the rock weak. Removal of load from *A* disturbs balance; pressure from surrounding parts of the crust—especially from a loaded segment such as *B*—starts slow flowage in the weak zone at depth, forcing the lightened segment *A* to rise until equilibrium is restored. If the rock forced up under *A* had the same density as the rock eroded from the top, uplift and erosion could continue indefinitely, with no permanent reduction in height of the highland. However, as explained on page 15, we have good reason to believe that the light granitic rock in a continent is underlain at moderate depth by heavier basaltic rock. If, as appears highly probable, flowage to restore balance occurs in the heavy "subcrust," highlands will be gradually lowered as erosion goes on, as shown by the following analysis.

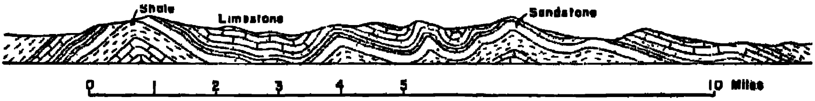
Granitic rocks have average specific gravity about 2.7. The dark-colored rocks are considerably heavier; at depths of some tens of miles the specific gravity may be as high as 3.3. As a given mass of rock is eroded from the top of block *A* of Fig. 6, the block must rise enough to admit an equal mass at its base, if balance with adjacent blocks is to be maintained. However, dense rock rises at the base to replace lighter rock removed from the top. Equal masses of these rocks will have volumes inversely proportional to their specific gravities. That is, a layer of the deep-seated rock 2.7 feet thick is equal in weight to a layer of the rock at the surface 3.3 feet thick. Since $3.3:2.7 = 11:9$, 900 feet thickness of the dense rock rising at the base of the block will exactly offset the loss of 1100 feet thickness at the top. Therefore 1100 feet must be eroded from the highland in order to effect a net lowering to the extent of 200 feet. If the highland has an initial average height of 4500 feet, to lower it by 4000 feet will require removal through erosion of $(4000/200) \times 1100$, or 22,000 feet of rock. Thus a mountain chain is a remarkably persistent feature; it can be reduced to a low hill-country only by continued erosion through a vast span of time. Nevertheless, there are many old mountain belts in various stages of destruction, including complete leveling by the forces of erosion (Chap. 19).

A. W. ROGERS.

Fig. 7. Folded strata along the Buffels River, north side of the Cape Ranges, Cape Province, South Africa. The folds are overturned toward the right (north).

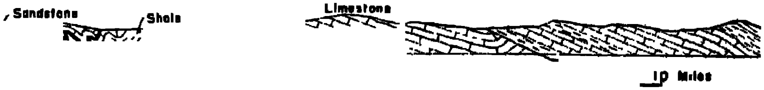
Renewed Thickening of Continental Masses. Material eroded from highlands is carried by streams and dumped into the sea at continental margins, building out deltas, coastal plains, and continental shelves. Thus the continents grow wider, and, as the areas planed down by erosion are repeatedly uplifted to maintain isostatic balance, the continental plates made of light granitic rock tend to become thinner. If no force intervened to offset this tendency, all highlands would eventually disappear, erosion by streams would slowly wear the continents almost to sealevel, and the waves of the sea would erode inland until all the lands were consumed. Actually, although the Earth is very old, mountains and plateaus stand at great heights, and some youthful mountain chains, as in the East Indies and coastal Alaska, are actively growing today. Many of the greatest highlands, including the lofty Mount Everest and the Tibetan Plateau, bear on their summits strata that were formed in shallow seas. Clearly, therefore, something is at work to defeat the slow leveling action of erosion. What forces are involved in thus renewing the lands, and how do they operate?

In all great mountain chains, those old and much eroded as well as the lofty ranges of more recent date, sedimentary strata have been strongly buckled and crumpled (Figs. 7, 8), and rocks of all kinds have been riven into great slices which are crowded together, one on another, overlapping like shingles on a roof (Fig. 9). This kind of deformation indicates that powerful compression has acted horizontally, driving masses of rock together and heaping them up in the long mountain belts. However, we have considered the evidence (p. 16) that a great mountain mass can rise above the general level only because it is buoyed up by light-weight rock extending to exceptional depth (Fig. 4). Therefore horizontal compression, in making mountain chains, has not simply forced great wrinkles above the general level of the Earth's surface. The plate of continental rock in the mountain zone must have been thickened by the powerful squeezing, and the mountain masses now reflect the zone of thickening, above which they are held in isostatic equilibrium. Such a newly formed belt of highlands is attacked with special vigor by the agents of erosion, and the resulting debris, consisting of boulders, pebbles, sand, and mud, is carried by streams to the lowlands and into the sea. As weight is thus removed from the highland, uplift occurs repeatedly because of flowage at depth to maintain isostatic balance (Fig. 6). After a vast span of time the thick-



U. S. Geological Survey.

FIG. 8. Layers of limestone, shale, and sandstone, deposited on the floor of a former sea, and later strongly buckled by compressive forces. Appalachian Mountains, western Virginia.



U. S. Geological Survey.

FIG. 9. Layers of limestone, shale, and sandstone, originally horizontal, now in great overlapping blocks that were driven together by compressive forces. Appalachian Mountains, southwestern Virginia.

ened part of the continental plate grows thinner through continued erosion, and the former mountain belt becomes a low, hilly country, although evidence of the earlier history remains in the buckled and fractured bedrock. Thus we recognize the "stumps" of ancient mountains in parts of France, Great Britain, and southern New England where no high topography exists at present. Mountain belts in various stages of destruction testify to repeated action, through geologic time, of the forces that thicken the continental plates and thus offset the leveling effects of erosion.

Large Unsolved Problems. Thus by the scientific method, using the observations and reasoning of many workers through several generations, geology has arrived at a general comprehension of the mechanism that has governed the changing face of the Earth through its long history. This is not to say that all the mysteries, or even those most profound, have been solved. We can only speculate on how the continental masses first came into existence as plates of light-weight rock covering about a third of the solid body of the Earth. Moreover, we do not know the nature and origin of the forces that have repeatedly deformed and thickened the continental plates. Speculations on some of these challenging problems are outlined in later pages (Chap. 19). This speculative aspect of geology is fascinating and valuable, although sometimes dangerous to the uninitiated. Use of the scientific method must take account of speculation along with fact and logical inference, distinguishing carefully among them. Careless broadcasting of hypotheses as if they were demonstrated truths has given rise to unfortunate misconceptions about the Earth and its history.

CONTEST BETWEEN INTERNAL AND EXTERNAL FORCES

From this brief analysis, it is clear that agents of two distinct kinds act on and modify the outer part of the Earth. The agents of one kind attack the rocks exposed at the surface, breaking them to pieces, changing them chemically, and moving the resulting debris to lower levels. Air and water, under the control of gravity, are the chief agents concerned in these activities; the essential energy for their work is supplied by the heat and light of the Sun. As explained in detail on later pages (Chap. 3), this energy is expressed in complex physical and chemical processes by which rock is *weathered* into soil and other loose material. Solar energy also causes evaporation, from seas and land surfaces, of the water which later is precipitated as rain and snow. The resulting streams and glaciers are the most powerful eroding agents, carrying away rock debris already formed and cutting directly into the bedrock (Chaps. 5, 9). Waves and currents of the sea, which also are manifestations of solar energy, erode the margins of the lands and spread out on the sea floors the waste brought down by streams (Chap. 11). All these agents combine in *grading* the Earth's surface, cutting down elevations and filling depressions; hence they are commonly known as the agents of *gradation*. If they could work without interruption, they would in time cut all the lands below sealevel and carry the loose products into deep water, below the effective reach of waves and currents (Chaps. 6, 11).

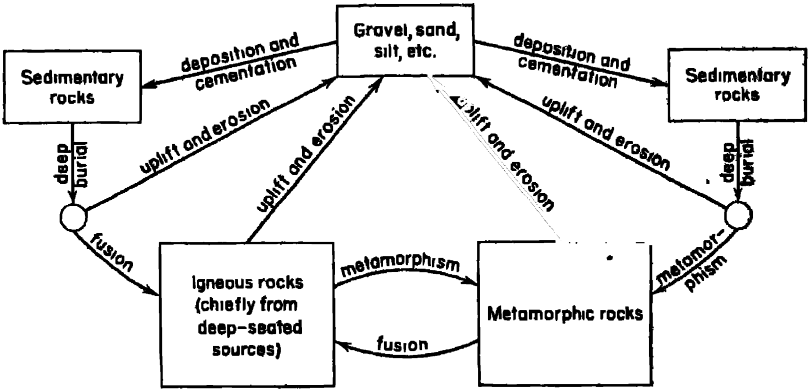


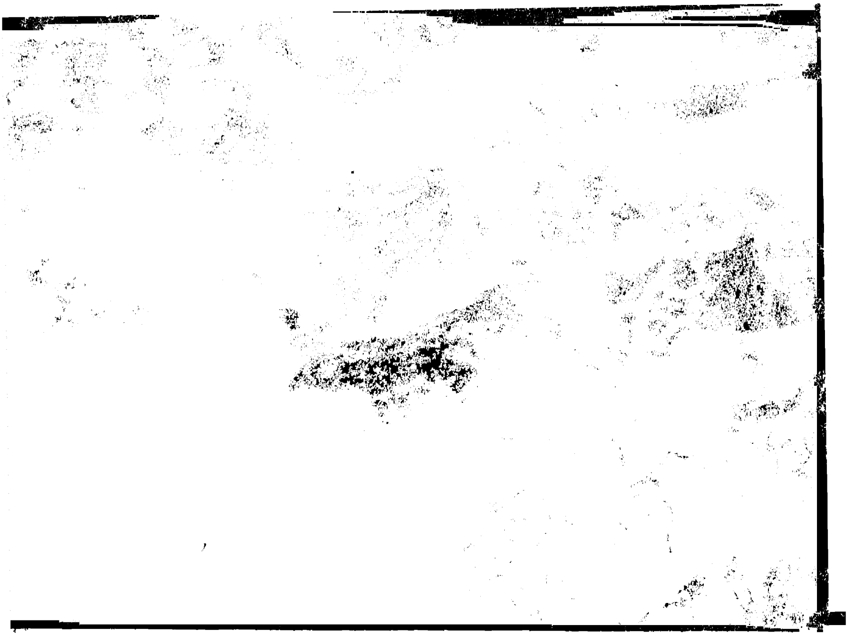
FIG. 10. Diagrammatic representation of "rock cycles," indicating that material in the Earth's crust may pass through the igneous, sedimentary, and metamorphic states repeatedly and in several possible sequences.

Since marine sedimentary rocks underlie large areas of the present continents, even in the highest mountains and plateaus, manifestly there are forces within the Earth that cause widespread uplift and thus tend to offset the wearing away of the lands by destructive surface agents. These internal forces elevate and depress large areas by gentle bending or *warping* of the crust, a process that has repeatedly caused emergence and submergence of large land masses in all the continents. In part these warping movements result from isostatic adjustment to loading and unloading, as material is shifted from one area of the Earth's surface to another (pp. 18, 19). However, such a mechanism of adjustment does not explain why areas that were for long periods subsiding sea floors have later been warped up to form high plateaus. The forces within the Earth also cause violent crumpling and breaking of the rocks in mountain zones, which rise as great welts above their surroundings. Volcanic products, rising from pockets of molten rock produced by localized heat, build up high peaks and extensive plateaus. Thus there are internal agents co-operating to raise the lands and increase their irregularity, in contrast and in opposition to the grading effects of the agents controlled by solar energy.

In the agelong contest between the internal and external Earth forces, much of the rock material in the crust passes through a succession of major changes. For example, granite exposed in a mountain region decays and breaks up, and the resulting debris is carried by streams into the sea, where it is built up, layer by layer, into a thick section of sediments. Perhaps at a later date these sedimentary layers are caught in the powerful vise of mountain-making forces, which not only crush and elevate the rock mass but also, as a result of great and long-continued pressure, change the mineral composition and appearance of the layers until they are no longer recognizable as sedimentary rocks and are classed as metamorphic rocks. As this mountain mass is attacked by the weather and by running water, the rocks are again reduced to gravel, sand, and smaller particles, which are swept away to be spread out in a new series of sedimentary layers Some of the rock material now visible at the Earth's surface may have passed through several of these major cycles of change, each requiring many millions of years (Fig. 10).

The Time Factor in Geology

From the geologist's viewpoint, therefore, the Earth is dynamic and changing; not static and inert. However, since most of the changes are slow as judged by human standards, it is necessary to form some conception of geologic time in order to appreciate the continuity of Earth history. The study of geology revolutionizes ordinary notions of time, just as a moderate knowledge of astronomy gives a new vision of space. Our Solar System has grand dimensions, and yet in its entirety it is but a point by comparison with the stupendous diameter of the starry galaxy. Similarly we think of the earliest human records as very ancient; but in a geologic sense the first appearance of man, some hundreds of thousands of years ago, is a modern event. The oldest relics of primitive men are found only in the superficial rock debris formed in the latest geologic epochs. In the thousands of years that have elapsed since the oldest civilizations existed, the major landscape features of the Earth have remained essentially unchanged; but we know from the immensely longer geologic record that, before man appeared,



KOLB BROS , GRAND CANYON, ARIZONA

Fig 11. Grand Canyon of the Colorado The inner gorge (of which only a limited segment appears in the view) is cut into granite and metamorphic rocks; it is about 1000 feet deep. The higher part of the valley is cut through flat-lying layers of sandstone, shale, and limestone, most of which were formed on ancient sea floors The steep cliffs are developed on resistant layers, and the gentler slopes on weak layers The outer rim is several miles distant from the river and 5000 to 6000 feet above it

generations of mountains were made and worn away by the same deliberate forces now at work. The length of time required for such transformations has been almost inconceivably great. How long has it taken for the Colorado River to excavate the Grand Canyon? We are astonished at the realization that the slow action of water has carried away so many cubic miles of solid rock. Nevertheless the Grand Canyon is a youthful feature, in the perspective of Earth history. The general record of earlier events is written plainly in the rocks (Fig. 11) traversed by the trail in climbing the vertical mile from the inner gorge to the outer rim Mountains were made and worn down; then the land was submerged beneath the sea for long ages and continued to sink slowly while mud and sand built up deposits thousands of feet thick; this loose material was converted into firm rock; and finally the rising of the wide plateau region high above the sea, in late geologic time, permitted the cutting of the canyon.

Since the whole of written human history is comprised within a few thousands of years, it is difficult to accustom ourselves to the much greater time scale of Earth history. The statement that some rocks are more than 1000 million years old excites an understandable skepticism. However, this statement is the result of exact calculations based on dependable evidence found within the rocks themselves Uranium and other radioactive chemical elements disintegrate continuously, at a measurable rate which seems to be absolutely uniform under all known conditions of temperature and pressure. As soon as a mineral containing uranium has been formed in a rock, helium gas and other products that result from the breaking down of uranium begin to be differentiated. The end product of the change is one kind of lead, which accumulates in the mineral. The ratio of this lead to the uranium steadily increases with the passage of time; and, since the rate of the change has been determined precisely, the age of a uranium mineral can be calculated after a chemical analysis reveals the exact ratio of lead to uranium in the mineral. It is as if a clock had been set going in each of these minerals at the moment of its origin; the geologist, with the aid of the chemist and the physicist, is now able to read the exact time on each of the clocks. By means of this remarkable method it has been determined that the length of geologic time is at least 2000 million years (Appendix D). In such a vast span of time it has been possible for the slow processes now in operation to fashion and destroy generations of mountains and to accomplish other astonishing changes to which the rocks bear eloquent testimony. The closest study of the geologic record has revealed "no vestige of a beginning, no prospect of an end."

Because most of the changes in the surface forms of the Earth and in the structure of the bedrock proceed slowly and continue through long ages, obviously much of the evidence by which we decipher the full history has to be circumstantial rather than direct. Some geologic processes, such as the decay of rocks at the surface, erosion by streams and by glaciers, and volcanism, can be seen in operation. The effects of these processes, though small during the lifetime of a man, clearly are cumulative. No observer doubts that the vast pile of Mount Etna has resulted from repeated eruptions like those that have been witnessed by living persons, though only a small part of the growth of the mountain has occurred during recorded human history. Similarly the full effects of stream erosion in fashioning landscapes come to be appreciated through comparative study of numerous stream valleys which represent obvious stages of development, from youthful to very old.

The introductory study of physical geology examines the processes now operating on and within the Earth and considers the evidence that all the visible features, in surface forms and in the bedrock, have been produced by these same processes, during geologic time. In other words, the chief objective in the study of physical geology is to make clear the methods by which the geologic record may be deciphered.

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CHAPTER 3

WEATHERING AND ITS PART IN EROSION

Bedrock and Mantle. Bedrock underlies all parts of the land surface, and, although it is concealed in large areas, it can be reached by digging or drilling to sufficient depth. The cover, which ranges in thickness from a few inches to hundreds of feet, consists of soil, clay, sand, gravel, and other loose material. Obviously the coarser pieces are fragments of rock or of individual minerals; and a powerful microscope shows that rock particles, some fresh and others very much altered, make up a large part of the fine material also. This complex assemblage of rock debris that covers the bedrock is the *mantle* (Fig. 12).

Careful examination of the mantle in a wide region reveals that in some places all the recognizable particles consist of minerals found in the bedrock directly beneath. In other places, however, most or all of the constituents have no apparent relation to the underlying rock. It is a logical inference that the mineral fragments in the mantle have been detached from firm bedrock and that much of the debris has been removed from its place of origin. The part that remains essentially in place is *residual* mantle; the part that is carried away and dropped elsewhere forms *transported* mantle.

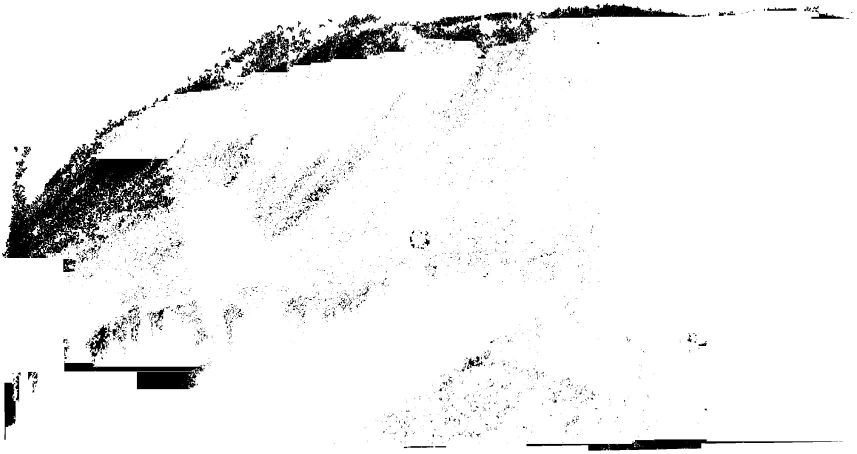
Weathering and Erosion. Wherever the bedrock is exposed at the surface, as in a cliff, it has been obviously changed. Usually some blocks or chips of the rock are loose, and others can be detached with little effort. Many of these pieces are so rotten that they crumble in the fingers. By quarrying into the cliff we encounter firmer, fresher rock; but along joints or crevices some staining and softening are evident to considerable depths. Even the most casual observer recognizes that the outer part of the rock has been changed by exposure to the weather. Accordingly we say that the rock has been *weathered*, and the rather complex set of processes involved in the breaking up and decay is called *weathering*. Aside from its fundamental position among geologic processes, weathering is of vital importance to the human race because it results in soils which are the basis of all life on the lands.

Most cliffs also suggest plainly that they are remnants left by the disappearance of great quantities of rock around them (Figs. 11 and 35). Regular layers end abruptly at a cliff face, and since identical layers can be recognized in neighboring cliffs or hills we can hardly avoid the inference that the layers once were continuous across the intervening space. It is difficult to imagine the removal of so much resistant rock; but the weathered condition of the rock now exposed at the surface suggests that this process of decay, continued for a very long time, has greatly facilitated the work of running water and other agents in fashioning the present landscapes.

The loosening and removal of rock material by any process at the Earth's surface is *erosion.* Since streams, glaciers, the sea, and the wind are important agents involved in actually removing the debris, we recognize as separate processes *stream erosion, glacial erosion, marine erosion,* and *wind erosion.* In each of these processes considerable fresh rock is carried away in addition to more or less weathered material; but, both directly and indirectly, weathering increases enormously the rate and effectiveness of the other erosive agents. Therefore the weathering of rocks is one of the most important parts of general erosion.

HOMER P. LITTLE.

Fig. 12. Piedre Mountain, near Creede, Colorado. The loose rubble on the slopes in the background is as coarse as that in the foreground. Some of the bedrock shows in prominent outcrops on the mountain side.



WEATHERING

Role of the Atmosphere. The chief agent of weathering, as of the weather, is the atmosphere, which uses energy derived from the heat of the Sun. Although it is for the most part invisible, the atmosphere is responsible for vast changes on the Earth. By carrying moisture from the sea and precipitating it on the continents the atmosphere makes possible the formation of streams and glaciers, which are powerful eroding agents. Winds, by blowing dust and sand, accomplish some erosion directly, and by generating waves on the sea they cause marine erosion of the coasts. Therefore the greater part of erosion is chargeable either indirectly or directly to the atmosphere. For the present, however, we shall consider only its role in rock weathering.

The important geologic activities of the atmosphere depend on its peculiar physical and chemical properties. Since it consists of gases it penetrates readily into all crevices and other openings that lead down from the Earth's surface and so comes into contact with much of the bedrock in addition to exposed surfaces. About three-fourths of the atmosphere consists of the gas nitrogen, which is rather inert chemically, although with the aid of certain plants and bacteria it enters into some important reactions in the ground. Oxygen, which makes up most of the remaining one-fourth, is enormously more active chemically and plays a large part in the attack on rocks. Water vapor is a highly variable constituent of the air, but its geologic role is large. Precipitated as rain or dew, it moistens the rocks, dissolves some of the minerals directly, takes an essential part in many chemical reactions, and assists in mechanical changes also. Carbon dioxide, which normally forms only 3 parts in 10,000 of the atmosphere by volume, has an importance out of all proportion to its rank in quantity. Aside from its vital function in supplying carbon to growing plants, carbon dioxide combines with certain of the elements in rocks and forms new mineral substances, some of which are readily soluble in water.

Other substances form small fractions of the atmosphere, but oxygen, water vapor, and carbon dioxide are the only constituents that are of large direct importance in rock weathering.

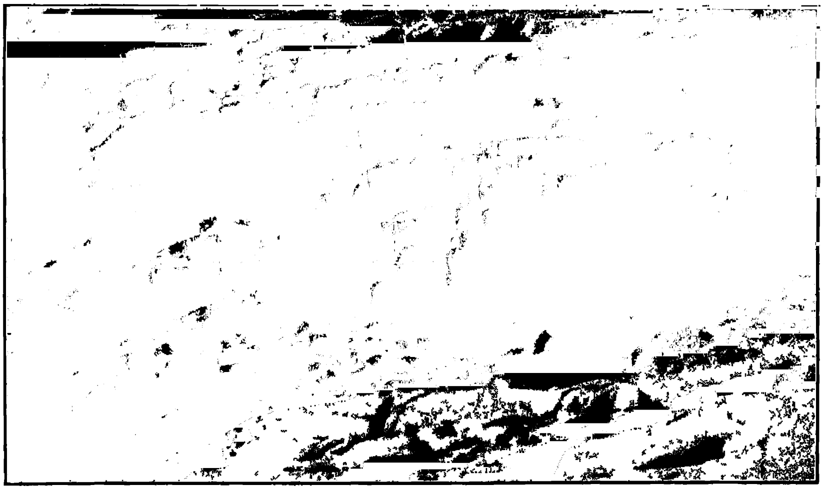
Other Factors Involved. Temperature plays a significant part in weathering. High temperature speeds up chemical action, and low temperature causes disruption by freezing.

Plants and animals help directly in the disruption of rocks and indirectly but more effectively by producing substances that attack minerals chemically.

Water, after it is precipitated and is no longer part of the atmosphere, continues to be a prime factor in weathering as long as it remains on the surface or at shallow depths in the crevices of bedrock and in the pore spaces of the mantle.

Gravity is a local factor in weathering, since rock masses are shattered by falling from cliffs or sliding on steep slopes.

Mechanical and Chemical Weathering. A twofold division is recognizable in the general effects of weathering. On the one hand, whatever the original nature of the rock, it is broken into smaller pieces. This mechanical breakup of rocks, with or without some chemical alteration, is called *disintegration.* On the other hand, the composition of the constituent minerals is changed by chemical attack, resulting in *decomposition* or "rotting" of the rock. Disintegration and decomposition usually proceed together and are so interrelated that they can not be distinguished clearly as results of wholly different processes. However, under certain conditions one effect predominates over the other. Disintegration and decomposition of bedrock are facilitated by the presence in nearly all rocks of numerous cracks and crevices (Fig. 13) which allow ready entrance of air, water, and other weathering agents. Even some of the mineral grains have cleavage planes or minute cracks which admit some moisture. By weathering processes these original



G K Gilbert, U. S Geological Survey.

Fig. 13. Fractures in the bedrock which aid weathering by allowing ready entrance of water, air, and plant roots. Sierra Nevada, California

openings are widened and others are formed. The loose arrangement and high porosity of the mantle make it especially susceptible to continued weathering.

DECOMPOSITION OF ROCKS (ROCK ROTTING)

Mechanical breaking up of rocks makes decomposition much more effective, because the chemical agents can attack each newly made fragment from all sides simultaneously and can descend ever deeper into bedrock as new cracks are formed or old ones are extended. However, since decomposition is directly responsible for some of the mechanical disruption, it is advisable to consider the chemical aspect of weathering first.

Effects of Oxygen, Water, and Carbon Dioxide. A piece of bright new steel exposed to the weather becomes coated in a short time with yellowish brown rust. If the exposure is continued for months and years scales of rust fall from the surface of the steel, and finally the entire piece can be crumbled into brown dust. In popular parlance, the steel has rusted away or decomposed; in chemical terms, the iron of which the steel was chiefly composed has reacted with oxygen and water to form a new substance, hydrous iron oxide or *limonite.*[1] This is an example of *oxidation* (chemical union with oxygen) and of *hydration* (chemical union with water).

A piece of bright copper exposed to the air keeps its original appearance longer than steel but eventually turns green from union of copper with carbon dioxide in the air, to form basic copper carbonate. This reaction is *carbonation*, a process that attacks some substances in exposed bedrock, with consequent weakening of the rock

Another important chemical process related to hydration is *hydrolysis*, which is a consequence of partial dissociation of water during complex reactions that occur in the moist mantle. Hydrolysis plays a role in the weathering of feldspar (p. 33).

Rocks exposed at the Earth's surface consist of various minerals, and these react to the chemical agents of weathering differently and at very different rates. Iron in such minerals as black mica in a granite or pyroxene in a basalt unites slowly with oxygen and water to form limonite,[1] which makes a brown stain on the surface of the rock. But the reaction is far more complex than the last statement suggests. Before the iron in pyroxene, for example, is released to form limonite,[1] the entire original composition of the mineral is broken up, and then the various elements enter into new combinations, some involving carbon dioxide, oxygen, and water. The result is softening and partial or total disruption of the mineral grains, not only at the surface but also along any crevices into which moisture can penetrate. In fact, the change may be more effective at a slight depth than on a bare surface, since moisture lingers in the shelter of crevices and enables the chemical work there to continue without interruption, whereas intermittent drying retards decomposition on surfaces exposed directly to the air.

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[1] Used as field term for hydrous iron oxide, regardless of mineral identity.

Solution. Chemical reactions do not proceed readily between dry substances, as we know from laboratory experiments. Therefore one of the most essential parts played by water in the changes described above is to serve as a medium in which other reagents can work. Besides performing this essential indirect function, and in addition to its union with other substances in the process of hydration, water aids in decomposing the rocks by removing certain minerals in solution. Only a few common minerals are soluble in pure water, but many others dissolve in carbonic acid (H_2CO_3) which forms when carbon dioxide (CO_2) unites with water. For example, the mineral calcite (calcium carbonate, $CaCO_3$) is nearly insoluble in pure water, but carbonic acid converts it into calcium bicarbonate, $CaH_2(CO_3)_2$, which dissolves and is carried away by water percolating through the pores and crevices in the rock. Calcium carbonate is the chief constituent of limestone, and vast quantities of this rock have been carried away in solution (Chap. 7). Other less soluble mineral grains scattered through various rocks dissolve more slowly, but their loss weakens the rock and provides additional openings to be used by chemical reagents.

The continued removal of soluble material by water percolating through the mantle or through shattered bedrock is termed *leaching.*

Decomposition of Feldspar. Since feldspars are the most abundant rock-making minerals, their response to weathering merits particular attention. When feldspar decomposes, the chief product is clay, which is of great importance because it is one of the commonest materials in the mantle and also because it goes into the formation of shale, the most abundant type of sedimentary rock (p. 572). The chemical weathering of orthoclase, one mineral of the feldspar group, is outlined as follows:

$$\text{Orthoclase } (KAlSi_3O_8) + \text{water } (H_2O) + \text{carbon dioxide } (CO_2)$$
$$= \text{clay (various hydrous aluminum silicates)} + \text{silica } (SiO_2)$$
$$+ \text{potassium carbonate } (K_2CO_3)$$

Complete decomposition of feldspar requires considerable rainfall and a warm or temperate climate. The carbon dioxide involved in the reaction comes in part directly from the air but chiefly from decaying vegetation. Some of the potassium carbonate (potash) formed is dissolved by water and removed, but part of it is held by the clay and is used as plant food if the clay becomes transformed to soil (p. 46).

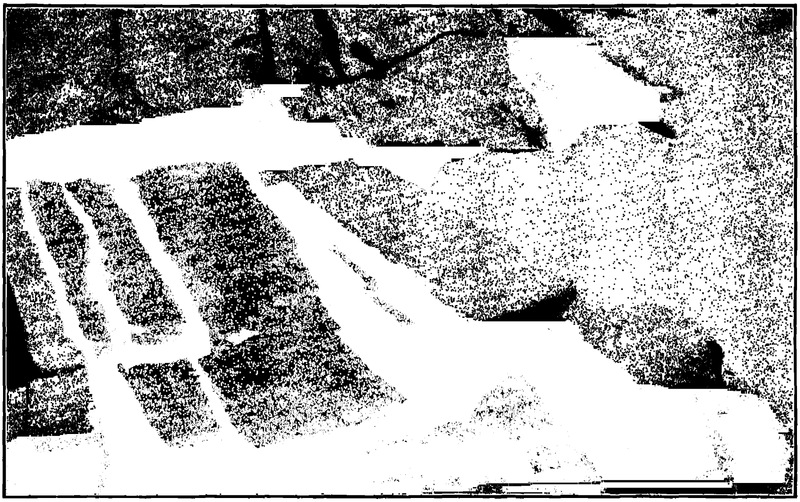
The generalized formula for decomposition of orthoclase may suggest that the process is simple and altogether inorganic. Nothing could be farther from fact. Reactions that take place in the decomposing mantle are so complex that even after long and concentrated study by soils specialists the exact nature of these reactions is in large part obscure. It is realized that an important part of the change is *biogenic;* that is, brought about through the agency of organisms, both plant and animal. Moreover the products of these reactions are extremely complex. Clays, the chief product of weathered feldspar, consist of several distinct minerals that exist in particles of submicroscopic size and therefore can not be studied by ordinary means. Use of powerful X-ray equipment and of the electron microscope reveals some of the properties of these minerals; but many important facts remain unknown in spite of the most ingenious modern techniques. Development and behavior of the clay minerals involve the difficult principles of colloidal chemistry. Many profound problems met in the study of clays must be solved before we shall have a reasonably full understanding of the soils that supply our food.

Chemical Weathering of Other Minerals. Amphibole, pyroxene, and other common minerals that contain aluminum also yield clay when they are decomposed. Several of these minerals contain some iron, which commonly becomes oxidized, and thus the clay acquires a yellowish or reddish color.

Quartz resists chemical weathering more successfully than any other common mineral in the rocks. Therefore, when feldspar and other minerals decay, the associated grains of quartz remain essentially unchanged. Since quartz is hard and has no cleavage to form planes of weakness, its grains also, better than those of most other minerals, resist mechanical wear by wind, streams, and waves, and therefore sand and gravel come to consist largely of quartz particles. Much of the sand that now lies on beaches or in shifting dunes is made up of quartz grains that were weathered from granites and related rocks during remote geologic periods.

DISINTEGRATION OF ROCKS (ROCK BREAKING)

Effects of Freezing Water. When water freezes it expands by nearly one-tenth of its volume, and if it is in a closed vessel the pressure on the walls is very great. In the days of muzzle-loading cannon this principle was utilized in disposing of captured enemy artillery. Rocks in exposed positions are disrupted by repeated freezing and thawing of water in cracks or pore spaces (Fig. 14); the mechanism is aptly called *frost wedging*. Although the water in most openings in rocks forms a system open to the air, preliminary freezing in the upper parts of water-filled crevices may form closed systems in which further freezing will cause bursting pressures. The effect is confined to a shallow zone and is most noticeable near the edges of cliffs where blocks are poorly supported. In high mountains, freezing at night occurs even during the summer months, and there the results are especially conspicuous. Probably the wedging apart of minute rock scales or mineral grains is as important in the long run as the dislodging of large blocks.



F. E. Matthes, U. S. Geological Survey.

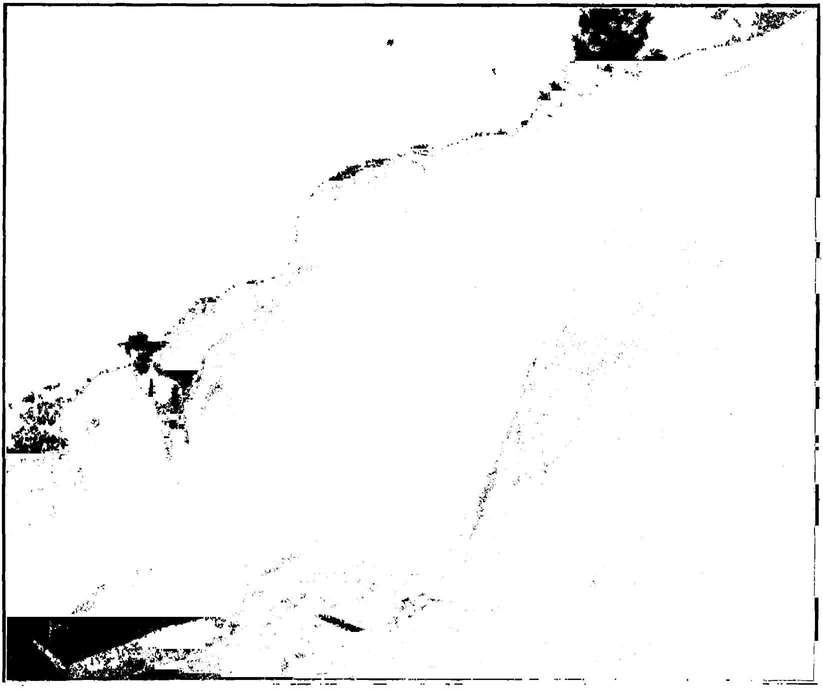
FIG. 14. Blocks of granite sprung apart by frost wedging. Sierra Nevada, California.

Direct Effects of Temperature Changes. The heat of a forest fire causes large flakes and spalls to break from exposed surfaces of granite and other rocks. Unlike iron and most other metals, rock is a poor conductor of heat; fire heats only a thin outer shell, which expands and is disrupted by the strains that result. Since forest fires set by lightning must have been common during long ages even before man came to disturb the economy of Nature, doubtless fire has been an agent of considerable importance in the disintegration of rocks at the Earth's surface.

Theoretically the smaller changes in temperature between day and night may also cause slow disintegration of rocks. Disintegration is conspicuous in deserts. The grains of unlike minerals in granite and similar rocks become loose and fall apart, and it has been argued that this effect is produced by minute stresses between mineral grains due to unequal expansion and contraction of feldspar, quartz, and other minerals when they are heated and cooled. The spalling off of thin flakes and slabs from rock surfaces, an effect that is more common in moist than in dry climates, also has been ascribed to expansion and contraction through repeated heating and cooling.

This theory has been tested by laboratory experiments, in which temperature ranges considerably greater than those recorded in any desert have failed to cause the slightest breaking of rock samples. In some experiments the changes in temperature have been repeated thousands of times by an automatic mechanism, to test the possibility that *fatigue* in the rock results from repetitions of small strains through a long period of time. Since no specimens subjected to such treatment have shown any changes detectable under powerful microscopes, there is serious doubt that temperature changes play any large part in weathering, except in connection with freezing water. It is probable that slight chemical changes, discussed below, are responsible for most of the effects that have been attributed to temperature changes.

Mechanical Effects of Chemical Weathering. Many chemical changes cause considerable increase of volume. Hydration especially produces swelling of the parts of the rock affected and therefore sets up strains. The outer surface of an exposed rock dries rapidly; but moisture that penetrates into minute crevices remains until some decay is effected. Slow increase of volume at slight depth finally disrupts the rock, and pieces that are quite or nearly undecomposed are broken off. *Exfoliation,* the spalling off of flakes and of comparatively thin concentric shells (Fig. 15), probably results chiefly from such chemical effects, although many observers have attempted to explain it as caused entirely by temperature changes. It is significant that exfoliation is common only in regions where moisture is fairly abundant, and not in deserts where rapid changes in temperature are most extreme. Detachment of thick shells of granite parallel to the surface of the ground is not a direct result of weathering. Many of the great shells that separate from the granite domes near Yosemite Valley, California, are 10 feet or more in thickness (Fig. 16); and it is highly improbable that either hydration or any other weathering process can be the original cause of the fracturing at such a depth below the surface. It has been suggested that relief of load by erosion of material from the top permits the rock to expand, with the result that thick shells are detached, as in the Yosemite domes and in Stone Mountain, Georgia. Exfoliated shells a fraction of an inch or at most a few inches thick are much more common, and these more logically can be attributed to hydration or other chemical changes, assisted by freezing of water that is admitted by the initial cracks.



G. K. Gilbert, U. S. Geological Survey

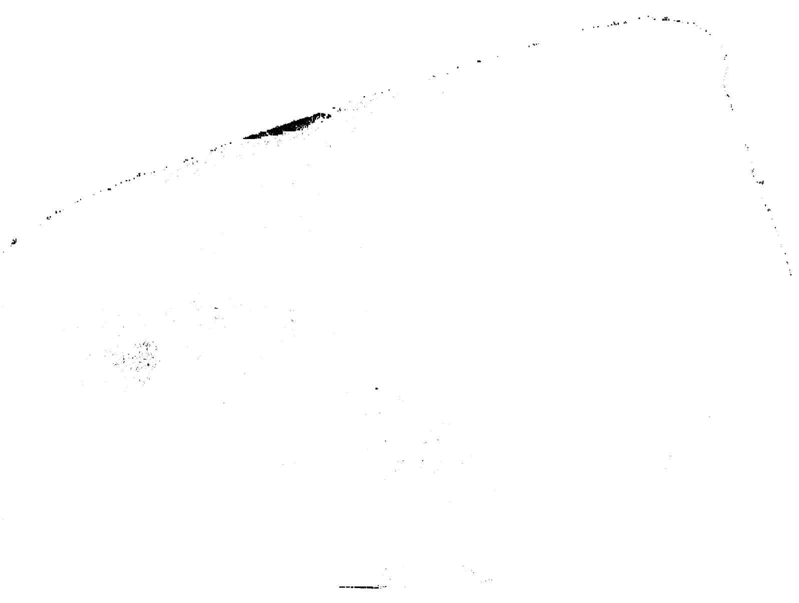
FIG. 15. Exfoliation of granite, near Nevada City, California. This locality is in a mountain district, where rainfall and temperature are moderate.

The slow disintegration of granite into its constituent mineral grains probably results at least in part from hydration in microscopic crevices, even in the driest regions. In Egypt the surficial disintegration of granite blocks and monuments is more pronounced in shaded places where moisture is conserved than it is on the sides most exposed to sunshine. Presumably the opposite would be true if temperature changes were the cause.

Mechanical Work of Plants. The wedging apart of rocks by the roots and trunks of growing trees has some importance in weathering. Shrubs and smaller plants also send their rootlets into tiny crevices of the bedrock and slowly enlarge the openings. In the course of time, the amount of disintegration accomplished in this way must be large, but much of it is obscured by chemical weathering which takes advantage of the openings as soon as they are formed, and therefore the effect can not be evaluated quantitatively.

FACTORS INFLUENCING THE RATE AND CHARACTER OF WEATHERING

Influence of Climate. Pronounced contrasts in climatic conditions are accountable for striking differences in the rate of weathering and in the effects produced. Since both warmth and abundant water are essential to effective chemical change in rocks, chemical weathering proceeds very slowly in polar regions even if moisture is plentiful, and in deserts even if temperature is high. Decomposition progresses most

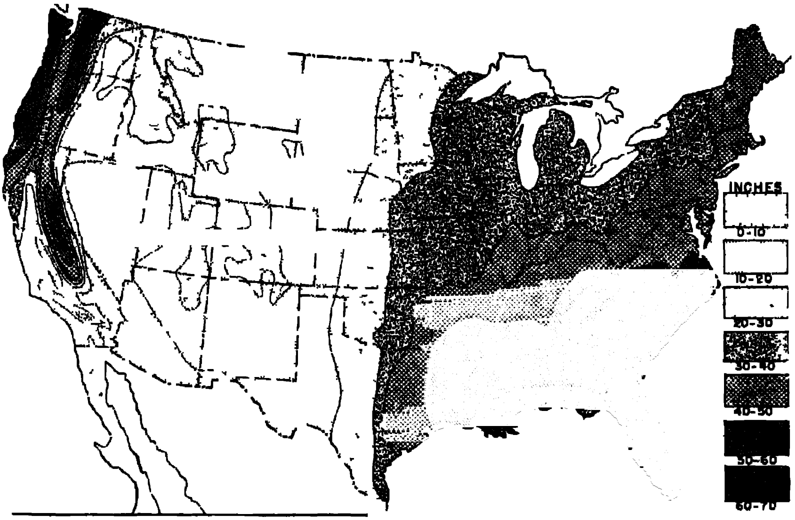


F C CALKINS, U S GEOLOGICAL SURVEY

Fig 16 Large shells of granite parallel to the surface The top of the cliff is 500 feet above the foreground, and some of the curved shells are 10 feet thick Northeast side of Half Dome, Yosemite Valley, California.

rapidly and reaches to greatest depth in moist tropical lands, although it is quite effective also in temperate regions that have abundant rain.

The controlling influence of rainfall appears convincingly in a comparison of the rainfall belts of the United States (Fig. 17). Decomposed mantle forms a thick cover in most parts of the southern Mississippi Valley and of the South Atlantic States. In the southern Appalachians, where rainfall is exceptionally heavy, the residual mantle attains thicknesses of tens and even hundreds of feet, and decomposition of its constituents is extreme. At the same latitude but with only one-fourth as much rainfall, the "panhandle" of Oklahoma has a thin mantle and numerous exposures of bedrock. In the arid region of southern Nevada and southeastern California, over great areas the bedrock actually appears at the surface or is veneered only with coarse rock debris that appears fresh or only slightly decayed. The coastal belt of Washington and Oregon has the heaviest rainfall in the United States, and the cover of decomposed mantle in that region is very thick, although the average temperature is considerably lower than in the southern Appalachians. In all moist regions of the temperate zones the luxuriant growth of vegetation is an indirect but powerful ally of the ordinary chemical processes. The dead plants build up in the



U. S. Geological Survey

Rainfall map of the United States. Conventional patterns show average annual rainfall in the several belts.

mantle a substance called *humus*, which furnishes carbon dioxide and thus increases the effectiveness of water in leaching the mantle and decomposing the bedrock. Arid regions are deficient in both soil and water for the growth of abundant vegetation, and therefore no humus is produced in such regions to reinforce the scanty water that is available for weathering.

It is evident that decomposition is dominant in moist parts of the country, and disintegration in the arid regions. It is not to be inferred, however, that forces causing mechanical breakdown are absent or of small consequence in areas where rainfall is large. Disintegration is very effective in such areas as long as bedrock is exposed; but the results are partly obscured by chemical decay, and, as a thick decomposed mantle develops, the bedrock beneath becomes more and more immune to mechanical disruption. In an arid region the forces causing disintegration may actually be less powerful, but since no insulating mantle of soil or clay is formed the work continues unabated and the cumulative results are conspicuously displayed. Frost weathering is limited in such regions owing to the scarcity of water, and there is no proof that any important disintegration results from direct heating and cooling of the rocks (p. 36). Plants play an appreciable part in disintegration only in areas sufficiently moist to support abundant vegetation. Volume increase from chemical effects, which probably is one of the most important factors in mechanical weathering, is far more effective in moist than in dry climates.

Frost wedging functions at its best in high-temperate latitudes or in corresponding climatic zones of high mountains, where alternate freezing and thawing occur during a considerable part of the year.

Influence of Topography. Since topography is an important element in determining climate, it exerts a large indirect effect on weathering. The average temperature decreases with increase of altitude, and hence frost action is effective in high mountains even in low latitudes Rainfall generally increases with increasing height, and therefore even in a very dry region the mountain peaks and ridges have sufficient moisture for some frost action and for considerable hydration of minerals, with resulting disintegration of the rocks by scaling or spalling, in addition to some decomposition.

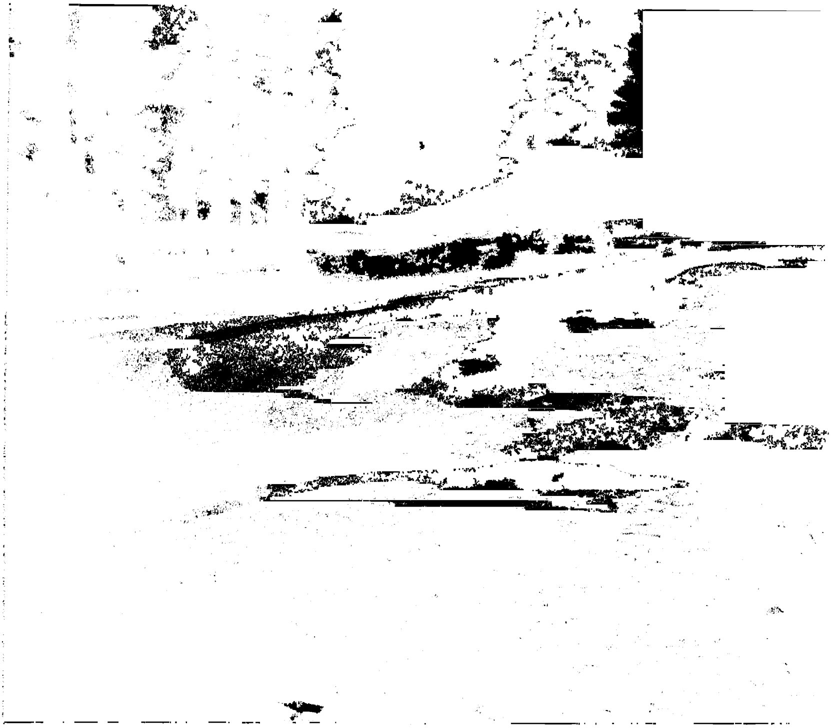
Topography has also a more direct control in the weathering process. Where slopes are steep most of the loosened debris falls, rolls, or is washed to lower levels, and thus fresh surfaces are exposed to the attack of the weather. At the bases of high cliffs the fallen blocks accumulate to form masses of *sliderock* (Fig. 12). Such deposits develop below cliffs in nearly all regions that have rugged relief, whatever the nature of the climate. Their presence is evidence of fairly rapid disintegration of the rocks above them.

Mountains in moist regions ordinarily have their lower, gentler slopes covered in large part with decomposed mantle. Rolling topography at a moderate elevation offers the most favorable opportunity for decomposition to extend to great depth. Vegetation protects the soil from rapid erosion, and as the upper part of the mantle decays more and more the chemical reagents penetrate to greater depth. There is a limit to this penetration, because below a certain depth all the openings are filled with water which moves very slowly or is almost stationary. Under areas that have moderate altitude, however, like some parts of the southern Appalachians, water in the ground drains downward, carrying carbon dioxide and organic acids to deeper zones and allowing the air to enter openings to considerable depth (Chap. 7).

Influence of Rock Composition. Rocks vary greatly in their susceptibility to weathering under different climatic conditions. Quartz is one of the most stable minerals, and rocks composed almost wholly of it, such as quartzite or siliceous sandstone, are especially stubborn in their resistance. Granite yields much more easily in any kind of climate; in a moist region the feldspar decomposes to form clay, and in a desert the grains of feldspar and quartz fall apart, probably through stresses set up by slight hydration of the feldspar. The actual rate at which these changes occur, however, is slow as judged by human standards. Granite has a long life in buildings and in monuments. In the Sierra Nevada some granite surfaces retain the polish and scratches formed by glacier ice thousands of years ago (Fig. 18).

Limestone is dissolved wherever there is sufficient water; caverns and sinks are formed (Chap. 7), and the limestone finally disappears, leaving only the small amount of clay, sand, or other insoluble impurities that may have been contained in it. In a moist climate, therefore, limestone weathers faster than many other rocks and tends to form lowlands; but in arid lands extensive solution can not occur, and there limestone is one of the most resistant kinds of rock. In many parts of Nevada and adjacent States, thick series of limestone layers make the highest mountain ridges and peaks.

Ordinary shale consists chiefly of clay which was the product of chemical weathering in an earlier geologic period. Further weathering can not cause much additional change in composition, but shale disintegrates easily into soft clay, especially if it becomes thoroughly soaked with water. In regions of old mountains like the Appalachians,



G. K. GILBERT, U. S. GEOLOGICAL SURVEY

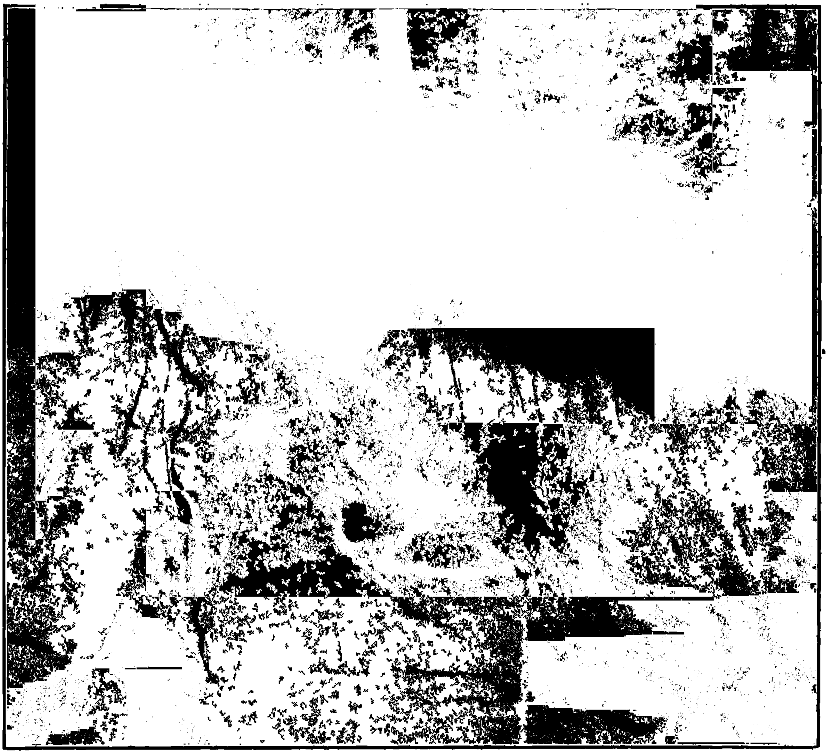
Fig 18 Granite surface high on the mountain slope above Yosemite Valley, California Movement of glacier ice during the latest ice age gave the bedrock a high polish. Although there may have been a covering of glacial debris that finally was removed by erosion, certainly the bedrock has been exposed for a long time; yet weathering has destroyed only part of the polished surface by loosening thin flakes and slabs of rock

where the rocks are steeply inclined sedimentary strata, the resistant sandstones come to form the high ridges, whereas the weaker shales and limestones underlie the valleys and plains Some sandstones, however, are not highly resistant. If the grains are cemented with calcium carbonate, solution removes the cementing material and the sand grains fall apart.

PRODUCTS OF WEATHERING

Residual Mantle. Wherever weathering occurs, part of the resulting debris is carried away by running water, wind, or some other transporting agent. In some places the loosened material is removed as fast as it is formed, and the bedrock is exposed as *outcrops* (Fig. 12)

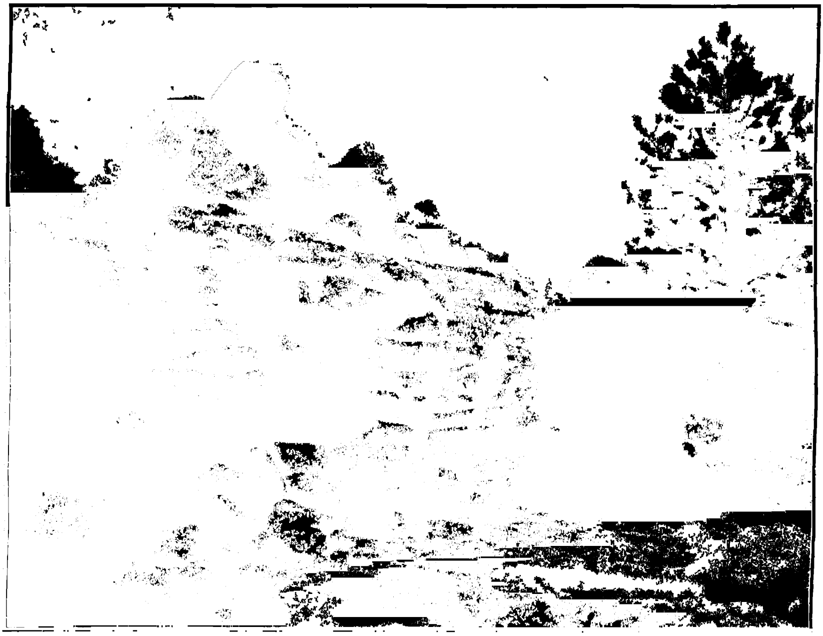
In many areas only part of the weathered rock is removed, by solution or otherwise, and a residue of the weathered material accumulates as residual mantle covering the bedrock to variable depth. In regions with plentiful rainfall the upper portion of this residual mantle normally consists of soil, and there is a gradual change downward through clay that contains some rock fragments into more or less rotten rock (Fig. 19). This evident relation to the underlying rock, shown by almost insensible gradation into it, is characteristic of mantle formed in place.



G. P. Merrill, U. S. National Museum

FIG. 19. Residual mantle on bedrock. The material grades from firm rock below, through rotten rock, into clay littered with rock fragments, and finally into soil made dark by humus. Near Washington, D. C.

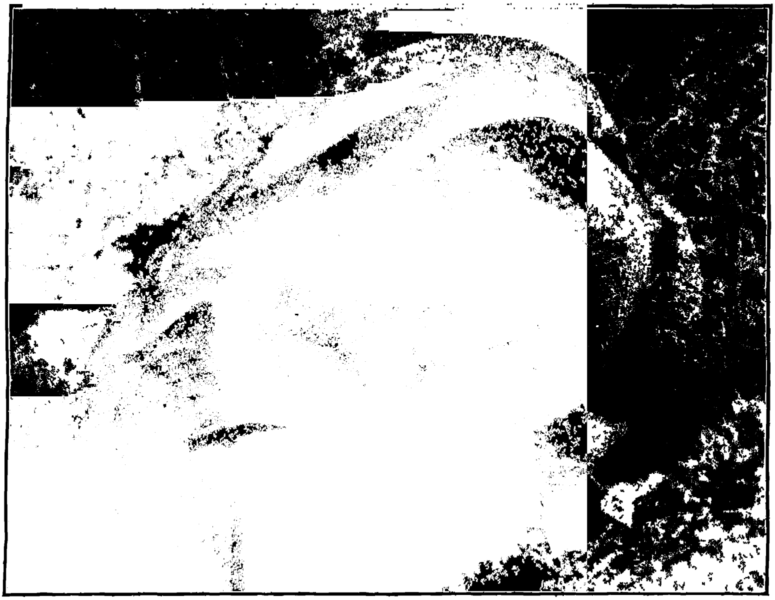
Residual mantle, of course, represents only a fraction of the original bedrock from which it was derived. Aside from continued losses that may have occurred through washing or blowing away of solid particles, all thoroughly weathered mantle has had a considerable percentage of its original mass dissolved and carried away by water. Such elements as sodium, potassium, and calcium, abundant in the feldspars of igneous rocks, go into solution when these minerals are decomposed. The salts contained in sea water, and in brines like those of Great Salt Lake and the Dead Sea, have come in large part from the leaching of weathered rock during long ages. The vast quantities of limestone now exposed on the continents represent calcium that was dissolved from older rock material and precipitated as carbonates on sea floors which later became land by elevation (Chap. 12).



W. T. Lee, U. S. Geological Survey.

FIG. 20. Residual boulders forming by weathering of jointed granite. Weathering progresses into each angular block, gradually rounding off the corners. Laramie Basin, Wyoming.

Residual Boulders; Spheroidal Weathering. In some situations most of the fine-grained products of weathering have been removed, but rounded boulders made of rock like the underlying bedrock are scattered on the surface. Many such boulders have been fashioned from rectangular blocks bounded by joints or cracks (Fig. 20). Since the corners of such blocks are attacked from all sides simultaneously they are rounded off, and continued exfoliation of thin shells finally makes the boulders nearly spherical or egg-shaped. Some of these *residual boulders* have an onion-like structure, with partly detached thin shells several layers deep (Fig. 21). This type of weathering is sometimes called *spheroidal weathering*. It takes place only in fairly moist climates, and since the minerals in the shells show the effects of hydration and other chemical changes, the belief that chemical effects rather than temperature changes are responsible for exfoliation is strongly supported (p. 37).



W. D. Johnson, U. S. Geological Survey.

FIG. 21. Granite boulder with concentric shells formed by weathering. Sierra Nevada, California.

Some bodies of rock contain masses of exceptionally resistant material which remain as residual boulders after weathering has reduced the weaker parts of the bedrock. For example, the chalk of southern England, well exposed in the cliffs of Dover, contains many nodules of dark flint. As the weak chalk is disintegrated and dissolved, the flint nodules are set free. Similar nodules of gray silica, known as *chert*, abound in some limestones in parts of the Appalachian region, the Middle West, and other sections of the United States. With the wasting away of the limestone the chert nodules accumulate until they litter the surface in large areas.

Soils. The term *soil* is sometimes applied erroneously to mantle of any kind, but properly it refers only to the part of the mantle which has been so decomposed and otherwise modified that it supports rooted plants. Development of a good soil is an extremely slow process. First the upper part of the mantle becomes sufficiently decomposed to yield some plant food, and a little vegetation takes root. Ordinarily the soil at this stage is scanty and poor in quality; it contains large quantities of disintegrated rock, little altered, and may be called an *immature* soil. As the plants die and are partially decayed they contribute some organic material, which contains carbon extracted from the air by the growing plants. Bacteria multiply in the soil, and some of these serve highly important functions such as taking nitrogen from the air and combining it with other elements to make available food for plants. The decaying vegetable matter furnishes acids that help to decompose the mineral particles in the mantle and leach out some substances, carrying them to lower levels. Thus the composition of the soil changes continuously, though slowly. If erosion does not disturb progress, and if there is plenty of rainfall to carry on chemical weathering and to support vegetation, the building of the soil accelerates for a time, since each step in the development increases plant growth and the resulting contribution of organic matter in turn speeds the development. Gradually the soil grows deeper, and as it loses all except the most insoluble of the original mineral particles it becomes *mature*. Usually the topsoil has a darker color than the underlying *subsoil*, which contains some materials carried down in solution but lacks the most vital ingredients developed in the topsoil. Generally a mature soil consists of several fairly definite layers, each from several inches to several feet thick, which are distinct in color and also in physical and chemical composition. The complete succession of these distinctive layers from the surface down to the unchanged parent material is called the *soil profile*. The following table represents, in somewhat simplified form, the profile of one soil developed in a moist tropical region at a high altitude.

PROFILE OF RESIDUAL SOIL ON GNEISS, TANGANYIKA, EAST AFRICA

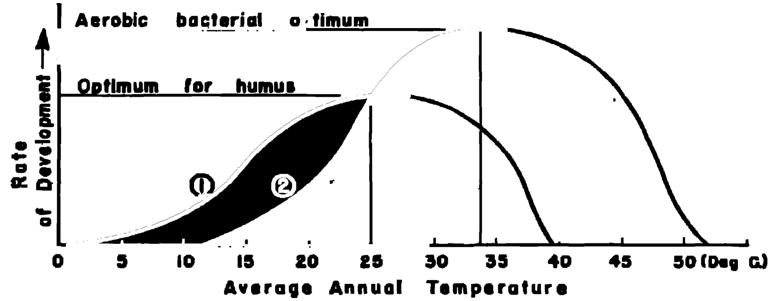
		% Clay	% Organic Carbon
0–8 in.	Dark, blackish brown topsoil	27.4	6.17
8–27 in.	Dark, orange-brown subsoil	40.6	1.72
27–70 in.	Mixture of clay with rock fragments		
Below 70 in.	Decomposing gneiss		

Most soil profiles contain a number of distinctive units, and many are highly complex. In a thorough study, the analyses give more constituents than the two shown in the table above. In particular, the clay is broken down into several fractions; most clays are complex mixtures, and the exact composition of a clay is significant to a student of soils.

As long as a soil developed on residual mantle is skeletal or immature it reflects the character of the underlying bedrock. An immature soil on granite (Fig. 19) is rich in clay formed from the weathered feldspar; it contains numerous bits of quartz and other constituent mineral grains of the granite, set free but unaltered; iron derived from biotite or from hornblende has been oxidized, and the resulting limonite colors the clay yellow; and the clay contains certain soluble substances such as potash which suggest the composition of the original feldspar. A soil derived partly from shale and partly from limestone contains an abundance of clay and is rich in calcium carbonate. However, as these unlike soils mature under similar climatic conditions they grow more alike, and eventually it is difficult to distinguish between them. In large areas of the United States the soils are more or less immature, and we commonly hear such terms as *granite soils* and *limestone soils* which suggest the recognizable relationship to the parent bedrock formations. There are only a few chief kinds of mature soil, and these kinds reflect the climatic conditions under which they were developed.

Soil Character Determined by Climate. To illustrate the effects of climate, let us consider the development of two strongly contrasted kinds of soil. In the cooler parts of the temperate zones, good soils are dark from their high content of humus, the organic matter derived from plants (Fig. 19). In tropical countries, on well-drained ground at low or moderate altitudes, soils contain no humus whatever, even if they support a dense growth of vegetation; humus in the low-altitude tropics exists only in swamps. The reason for this striking contrast between the tropical and the cool temperate zones is simple. Bacteria destroy humus in the soil; bacteria are numerous in all soils, but their development is retarded by the comparatively low temperatures in high latitudes, and therefore in those parts of the world humus develops faster than it is destroyed wherever plants grow abundantly. In the tropics the high average temperature stimulates bacterial growth to a maximum, and humus is destroyed as fast as it can form (Fig. 22). In swamps, however, aerobic bacteria can not thrive because the water excludes air from the soil, and therefore humus accumulates in tropical swamps. It is found also in high mountains in the tropics, where lower temperature checks the development of bacteria.

But the difference in humus content is not in itself the principal point in the contrast between typical soils in the two zones; the final consequences of the presence or absence of humus are very important indeed. With the aid of humus, water percolating through the soil gradually leaches out certain substances, especially iron oxide and aluminum oxide, depositing them at lower levels in the mantle. If no humus is present the water can not remove these substances, which therefore become concentrated in the soil as other materials are removed. Therefore the common soils of tropical countries are rich in alumina and in many regions are colored various shades of red by oxidized iron Such a soil is called *laterite*, from the Latin *later*, brick; the material has for centuries been cut and used as building bricks, as illustrated in the famous ruins of Cambodia in French Indo-China. In very old laterites these hydroxides are so concentrated and plant food is so deficient that plants grow poorly even with a high rainfall. Such soils cover large areas in India, Brazil, and other countries whose rainfall is controlled by the seasonal monsoon winds. Other mature tropical soils have a high content of alumina but lack the iron oxide coloring. Extreme development of alumina-rich soils has resulted in accumulations of *bauxite*, from which the metal aluminum is derived.



Modified from Parry Reiche.

FIG. 22. Curve 2 shows that aerobic bacteria (those that require free oxygen) thrive best in annual temperatures that average about 33.5°C. (92°F.). The rate at which plants contribute organic matter to soils increases up to 25°C. (77°F.) average annual temperature (curve 1), at higher temperatures, aerobic bacteria destroy all organic matter in well-drained soils.

Humus-laden soils are not devoid of iron and aluminum oxides, but the quantity of these substances continuously diminishes in such soils as they mature. The presence of some iron in nearly every soil is demonstrated by heating a sample to a high temperature in a kiln or oven; all the organic matter is burned out, and the iron compounds disseminated through the soil become *ferric oxide* (Fe_2O_3), which gives the whole mass a reddish color. Before the heating the iron was kept *reduced* by the organic matter to the *ferrous* condition, in which form the iron compounds have no conspicuous color. Mature soils in the warmer parts of the temperate zones are intermediate in character between those of more northern regions and the tropical laterites. In the southern Appalachians the thick soils on well-drained foothill slopes are distinctly red, indicating that bacterial action is vigorous in destroying humus and thus permitting accumulation of ferric oxide. However, in the stream valleys of the same region, where the mantle is more continuously wet and plant growth is more vigorous than on the slopes, the soil is prevailingly gray or black. Evidently in these situations the humus, if it does not actually leach out the iron, at least keeps it reduced to the ferrous state and so obscures it.

Soil Science. The study of soils is an important science in itself, quite apart from geology. It requires an intimate acquaintance with principles of chemistry and physics. The foregoing brief discussion is necessarily incomplete and oversimplified. For an adequate treatment the student must consult an extensive literature prepared by specialists.

DIFFERENTIAL WEATHERING

The unequal resistance of rocks to weathering is responsible for many details in landscape features under any climatic conditions, but particularly in arid and semiarid regions. In lands that have abundant rainfall the soil and other mantle conceal most of the bedrock and tend to obliterate small irregularities. Wherever there are cliffs or other outcrops, however, the rocks are sculptured into irregular forms by more rapid disintegration or decay in some parts of the rock than in others. This unequal yielding resulting from the nonuniform character of rocks is called *differential weathering.*

The scale of differential weathering has a wide range; the nonuniformity of bedrock is responsible for some of the largest features in a landscape and also for minute irregularities of microscopic size. Thick sandstone formations tend to make ridges while adjacent shales are worn down to form lowlands; great thicknesses of limestone disappear by solution under humid climates but show superior resistance by forming mountain ranges in arid regions. Of course other important factors besides weathering are involved in producing these major landscape effects. Most rocks that are susceptible to rapid weathering also yield easily to cutting by streams or other eroding agents, and therefore the final results have a complex origin. However, the part played by unequal weathering in producing uneven topography is fundamental.

The fashioning of many smaller features depends more directly on differential weathering effects. Alternating layers of sandstone and shale exposed in a cliff yield at different rates; gradually the shale gives way to form re-entrants between the projecting sandstone strata. If shale lies beneath thick sandstone an overhanging cliff results. Striking effects are produced from various combinations of resistant and weak beds, especially in semiarid regions. Jointed rocks are fashioned into irregular pillars, some of which have remarkable and impressive forms. There are numerous excellent examples in the Garden of the Gods, Colorado, and in Bryce Canyon, Utah. In many places large residual boulders are left delicately balanced, with only a small surface in contact with the rock beneath.

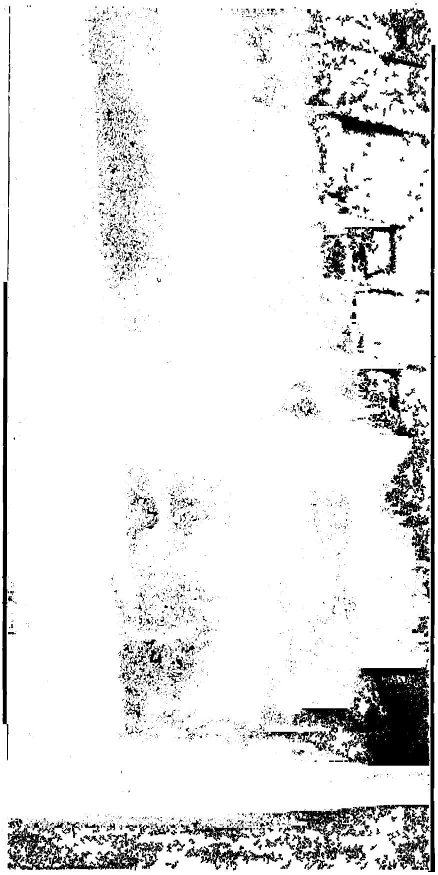
Quartz veins and small dikes of resistant igneous rock stand in relief as the surrounding rock is etched away. Veinlets that intersect in an intricate network form patterns that resemble lace or filigree work. Some sandstones that look quite uniform on fresh faces weather into a maze of pits and ridges that suggest honeycomb. This effect indicates irregular cementation of the sand grains.

Under certain conditions rock of uniform composition also weathers irregularly. Shallow depressions on exposed surfaces promote local decay by holding rain water and thus keeping small areas moist. As weathering proceeds the depressions deepen, and the effect is intensified.

PRACTICAL ASPECTS OF WEATHERING

Some appreciation of geologic time is necessary for a proper perspective of weathering processes and effects. Fresh rock shows little evidence of decay during the lifetime of a man. Ornamental building stone may lose its polish and become discolored; but centuries are required for the visible disintegration of resistant building blocks. Some old European castles bear eloquent testimony to the cumulative effects of attack by the weather (Fig. 23). Sturdy masonry of much more recent date requires frequent repairing, and we learn by inspection of stone structures in practically every city that it is poor economy to use weak, porous stone for constructing important buildings.

The soil is mankind's most priceless material heritage from past ages. Even after the bedrock has been broken up and decomposed, thousands of years are required for the development of fertile soil a few inches thick. Countless generations of plants must live and die, myriads of bacteria must run their life cycle, to supplement the physical and chemical weathering of rock material in the preparation of an agricultural soil. Once the humus-laden topsoil is swept away from a field, there is no practicable way to repair the damage quickly, even with the most laborious and expensive addition of prepared fertilizers. It is appalling to see the careless and ignorant waste of good soil in



C. R. Longwell.

FIG 23. Columns beside a doorway of Durham Castle, England, showing effects of weathering

large areas of this country. Recent widespread interest in soil conservation is a hopeful sign that a concept of the soil as a precious and perishable national asset is growing in the public consciousness.

RELATION OF WEATHERING TO OTHER ASPECTS OF EROSION

In the discussion above it has been necessary to point out repeatedly the interruption or modification of weathering effects by streams, the wind, or other agencies. Weathering is not an isolated process; it is intimately related to other activities at the Earth's surface, all of them striving toward the same end—the wearing away of the continents. Weathering attacks the rock in cliffs, and as tiny particles are loosened they are washed off by rain or blown away by the wind. Larger pieces fall to the ground, where weathering continues to reduce them while gravity and running water urge them down the slopes. Beneath the surface, weathering proceeds with less interruption and produces a considerable thickness of soil and other mantle; but as this cover forms it slowly migrates to lower levels by a process described later as *mass-wasting* (Chap. 4); continuously the water soaking through the mantle robs it by dissolving various substances; water at the surface forms gullies and washes the material into larger streams; and as the moving mantle arrives at the bottoms of valleys the streams take their toll directly (Chap. 5). Thus weathering and transportation of rock materials proceed together. Streams and the wind carry pebbles or sand grains along the ground, and by the friction other particles are worn from the bedrock or from the moving debris; but this process of frictional wear is distinct from weathering (Chaps. 5 and 10).

Some of the mantle in transport is dropped temporarily at the bases of mountains, on the floodplains of streams, on beaches, or in sand dunes; but in the upper parts of these deposits disintegration and chemical decay continue. There is no escape from the unrelenting attack until the debris is deeply buried on land or spread out in lakes or upon the sea floor, and there enters into the formation of new sedimentary strata (Chap. 12).

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These two papers indicate trends and methods of study in a highly specialized and difficult field of soil science.

CHAPTER 4

MASS-WASTING AT THE EARTH'S SURFACE

Weathering is essentially a static process, and any removal of weathered material from its place of origin is accomplished by other agents, of which the most conspicuous are rainwash, streams, glaciers, waves, and the wind. However, there are important movements of the mantle, for the most part slow but locally rapid and even catastrophic, which are not directly dependent on the common transporting agents and are closely related to the continued action of weathering. These movements, which are controlled directly by gravity, in large part escape notice because their average rate is imperceptibly slow; nevertheless they account for a large and essential part of erosion. Because the movements affect large masses of material as units, and because they are part of the general wasting to which all the lands are subject, the entire process including these movements is here designated *mass-wasting*.

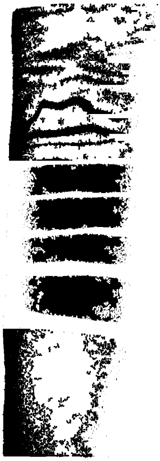
In some of its forms mass-wasting proceeds so slowly that it is detected only by circumstantial evidence. In other forms it is clearly perceptible, and exceptional movements take place at a rapid rate. There are all gradations between extremes in rate of movement. However, the processes involved can be grouped into those that operate gradually and those that are distinctly faster in their operation.

GRADUAL MOVEMENTS

Creep. The mantle on slopes, however gentle they may be, is subject to slow but continuous downward migration, known as *creep*. This movement goes on even though the slope has a close covering of grass and other vegetation that acts as a protective armor against the formation of gullies. As we look at such a slope it is difficult to believe that it loses anything to erosion except for some mineral matter taken into solution by water percolating beneath the surface; nevertheless the entire mantle is moving down the slope, carrying the cover of vegetation with it. The moving force is gravity. On even a moderate slope the weight of the loose material creates a tendency to slide or flow.

When the soil is soaked with rain, the water in the pore spaces adds to the weight and also lubricates the mass, thus decreasing the friction that retards motion. Gravity acts continuously, and even if movement is almost infinitesimal in the course of a year, the final result is inevitable.

Numerous factors contribute to the effectiveness of creep. In regions that have cold winters, water in the mantle tends to become segregated and frozen into layers of clear ice which separate layers of soil or clay (Fig. 24). Since the ice represents an addition of volume, there is uplift of the surface equal to the total thickness of the ice layers. This action is called *frost heaving*. On a hillside the surface of the ground is lifted essentially at right angles to a slope; when thawing occurs, each point tends to drop vertically instead of returning exactly to its former position (Fig. 25). Freezing causes expansion also parallel to the slope, with resultant displacement of particles chiefly downslope, under the influence of gravity. Animals that burrow in a slope pile most of the excavated material downslope; and later the burrows are filled by soil creeping from above. Plant roots in their growth wedge material chiefly downslope, and cavities formed by decay of roots are filled in the main by soil from the upslope side. Uprooted trees on hillsides tend to move downslope, urging soil and loose stones with them. As boulders and other rock fragments on a slope disintegrate, most of the detached particles fall or roll farther downslope. Slow removal of mineral matter in solution by percolating water creates voids in the mantle that tend to be filled by material urged down the slope. All these effects, individually small but all in the same direction, add up to large totals through countless repetitions.



Stephen Taber.

FIG. 24. A cylinder of clay, after freezing in an apparatus that was open with respect to water. The dark bands are ice, representing water that was drawn in continuously as freezing progressed. The top of the cylinder was elevated by an amount equal to the total thickness of the ice layers.

Although the rate of creep is slow as judged by human standards, in many situations it is at least roughly measurable. On one slope where a railway was repeatedly disturbed, the rate of creep was estimated as 6 to 10 feet in 50 years. Poles, fenceposts, and gravestones set on hillsides become tilted downslope, and in time require resetting. Stone and concrete retaining walls on such slopes are displaced or broken, and road grades become badly disturbed (Fig. 26). On some hillsides, tree trunks are tipped at noticeable angles; less commonly the trunks have a strongly convex bend downslope, the result of a conflict between creep and the tendency of trees to grow vertically. Even the upper part of weak bedrock is involved in the persistent creeping movement (Fig. 26).

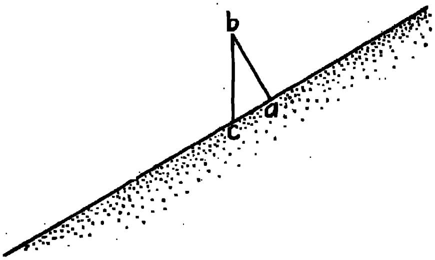
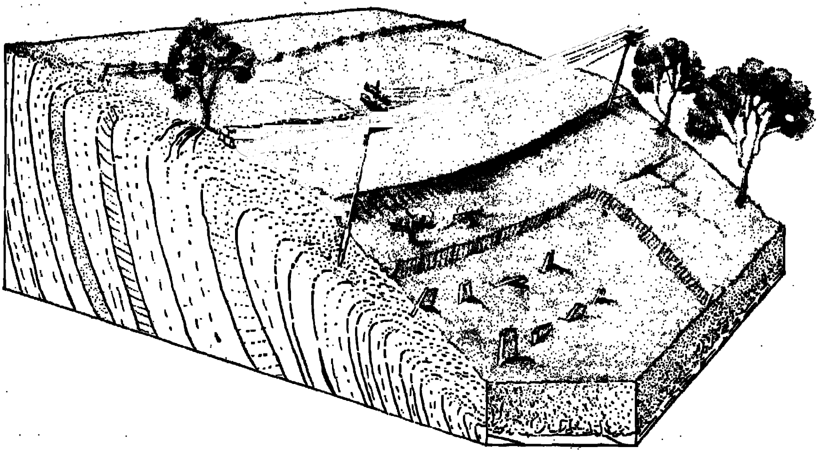


FIG. 25. When the soil on a hillside expands by freezing, surface material at *a* is carried outward, at right angles, to *b*; when thawing occurs and the surface subsides, the material tends to drop vertically to *c*. (Vertical scale exaggerated.)



Modified from C. F. Stewart Sharpe, Columbia University Press.

FIG. 26. Effects of creep on telephone poles, trees, walls and monuments, a road grade, and strata in bedrock.

A striking illustration of creep as affecting the works of man is furnished by Durham Castle, a stone structure built several centuries ago near the top of a steep slope in northern England. For a long time there have been conspicuous cracks in the walls, and a few years ago some of the cracks had grown so wide that partial collapse seemed to be threatened. On investigation it was learned that the foundations of the castle did not extend to bedrock but rested entirely in the thick mantle of the hillside. During the centuries this loose mantle has been creeping slowly down the slope and striving to carry the castle with it. As the rate of movement was not everywhere the same, the walls of the building were being twisted and ruptured. The danger has been removed, with considerable difficulty and at large expense, by strengthening the foundations and extending them to greater depth.

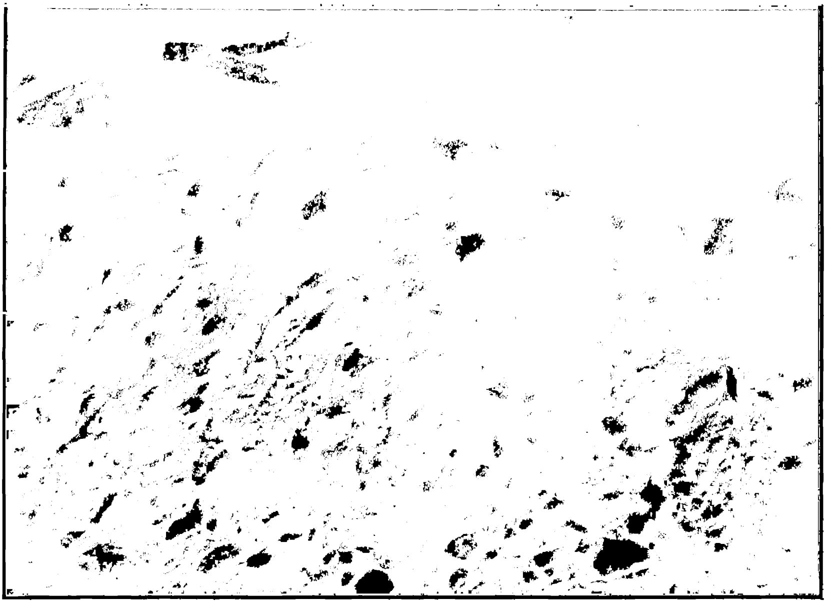

R. F. Flint.

FIG. 27. Unsorted solifluction debris exposed in road cut. South slope of Calispell Peak, northeastern Washington, at altitude 5000 feet. The section shown in the view is about 10 feet long.

Solifluction. A special form of creep is common in regions where the ground freezes to considerable depth. In warm seasons the upper part of the mantle thaws, while the lower part remains solidly frozen. Since the surplus water has no opportunity to drain downward, the upper unfrozen layer of soil, varying in thickness from a few inches to several feet, becomes saturated. Explorers in Arctic and high mountain regions find that mantle in this condition is like a viscous liquid and will not bear a man's weight. On slopes the saturated

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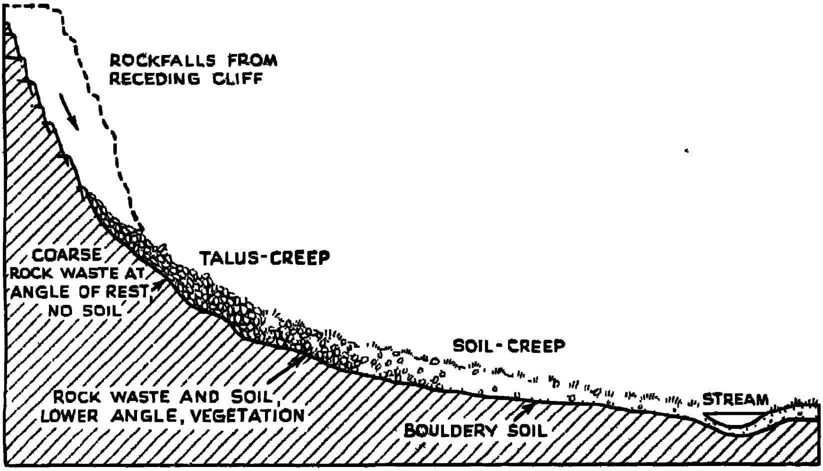
Fig. 28. A rock glacier in Silver Basin, Colorado. Slow movement of the rock debris is indicated by the pattern of surficial ridges.

mantle flows sluggishly, like a thick liquid; this type of movement is appropriately known as *solifluction* (soil flowage). One result is the mixing of soil with coarse rock fragments to form an unsorted mass of debris (Fig. 27).

Rock Glaciers. At the bases of exceptionally steep mountain slopes, rock debris accumulates rapidly, and in favorable locations masses of the debris take the form of rather definite "streams" that move slowly down slopes or narrow valleys. Freezing and thawing of water in such masses, together with gravity, produce gradual movement. From their general resemblance to valley glaciers (p. 165) these streams of rock fragments are called *rock glaciers*. They are especially numerous and well developed in Alaska and in the San Juan Mountains of southwestern Colorado (Fig. 28).

MORE RAPID MOVEMENTS

Sliderock and Talus. When rock fragments are loosened on the face of a disintegrating cliff they fall to the slope beneath, and by rolling or sliding reach a position of rest lower on the slope or at its base. Continued accumulation results in a heap or sheet of waste, called a *talus* (Fig. 29). This term applies to the body of rock fragments as a unit; the material composing the talus is *sliderock*. Typically the profile of a talus is a curve concave upward, steepening at the top to the maximum angle at which the coarsest of the sliderock will lie; for the largest rock fragments, this angle may approximate 40 degrees. In the upper part of a steep, growing talus the fragments lie in delicate equilibrium, so that the weight of a climber may precipitate violent sliding. There are frequent additions of new material, sometimes through the falling from the cliff of large rock masses that shatter from impact, but more commonly through the dropping of individual blocks loosened by frost action or the wedging of roots.


ROCKFALLS FROM RECEDING CLIFF
TALUS-CREEP
COARSE ROCK WASTE AT ANGLE OF REST, NO SOIL
SOIL-CREEP
ROCK WASTE AND SOIL, LOWER ANGLE, VEGETATION
STREAM
BOULDERY SOIL

From Sharpe, "Landslides and Related Phenomena," permission *Columbia University Press.*

FIG. 29. A talus in typical relation to bedrock and lower soil-covered slope. The creeping mantle is slowly carried away as it reaches the stream, at the right

Exceptional blocks that fall in a position favorable for rolling gain momentum as they descend and do not stop until they reach the lower gentle part of the slope. Generally, however, the freshest blocks lie in the upper part of the talus. In time they are disintegrated and decomposed, and the resulting fragments are carried down the slope. In a moist region the lower part or *toe* of a typical talus merges into a soil-covered slope set with grass or other vegetation (Fig. 29). Beneath such slopes there is slow creep of the mantle. The coarse debris in the talus also undergoes continuous creep (*talus-creep*), with occasional interruptions by local rapid sliding.

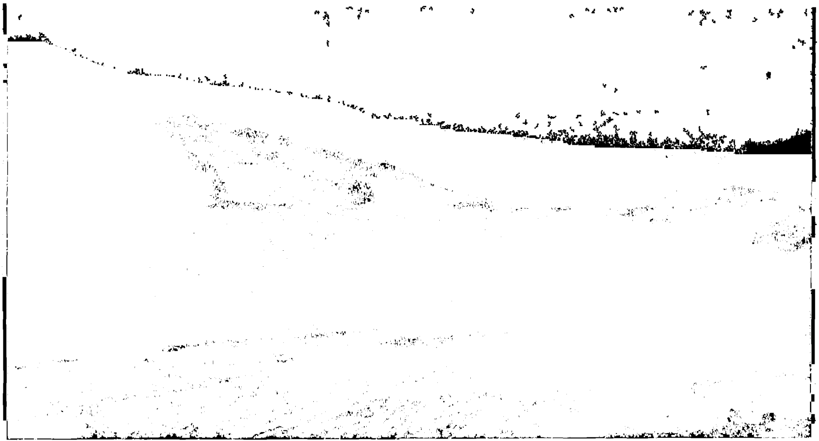
Many conditions affect talus development, and consequently there are wide variations in form. Vigorous streams that flow near some cliffs carry away all the finer debris, and only the large blocks remain to form a modified talus. On some mountain slopes the fragments dislodged from cliffs during the colder months slide or roll on snow of varying depth and build irregular heaps of debris known as *winter talus ridges,* which lie beyond the toe of the normal talus. Since frost wedging is particularly active in high mountains during late winter and spring, while the snow is deep and firm, winter talus ridges are strongly developed below some slopes that have little normal talus. After the snow has disappeared the ridges lie so far from the parent slopes as to seem puzzling unless the building process under winter conditions has been observed.

Landsliding and Related Processes. In regions with rugged topography, numerous large masses of earth and rock slide bodily down slopes, some of them abruptly and destructively, others so gradually that trees continue to grow on them almost undisturbed. Some of these *landslides* involve only loose material originally in the mantle; others include important masses of bedrock which, after displacement and more or less shattering in the sliding movement, become a part of the mantle.

One of the most famous slides in recent history occurred at Frank, Alberta, in 1903. A volume of rock estimated as about 40,000,000 cubic yards suddenly slipped from the top and flank of Turtle Mountain and rushed to the floor of the valley more than 3000 feet below. The rock was shattered by the impact of fall, and the mass of rubble spread in a sheet entirely across the valley, obliterating the town of Frank and killing 70 people. Momentum carried the front of the slide more than 2 miles out from the base of Turtle Mountain and 400 feet up the opposite side of the valley.

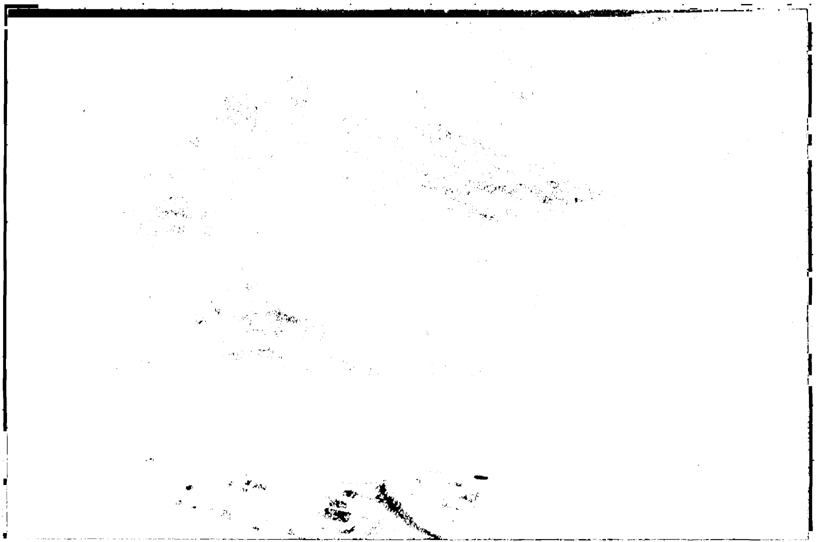
Although the Frank slide was an abrupt catastrophe, conditions that made it possible had been in preparation for a long time. The rock was weakened by a set of large joints inclined toward the valley. Water percolating along these fractures had slowly loosened parts of the rock by frost wedging and solution. Earthquakes that occurred two years before the slide probably caused further loosening, but not enough to start immediate movement. When the breaking point was finally reached, the inclined joints, lubricated with water, served as ideal surfaces for the initial slipping.

Other great landslides in the Rocky Mountain region have created large lakes by the damming of streams (Fig. 30). Disastrous landslides



C. R. Longwell.

FIG. 30. The 1925 Gros Ventre slide, as seen from a point about 2 miles upstream. The head of the slide, out of view at the left, is 2000 feet vertically above the valley floor. Momentum of the down-rushing mass carried its front far up the right side of the valley. The lake impounded by the slide had its highest surface lowered 60 feet as flood water cut a trough through part of the natural dam; the earlier high shoreline is best seen on the far side of the lake, at the left. Gros Ventre Valley, northwestern Wyoming

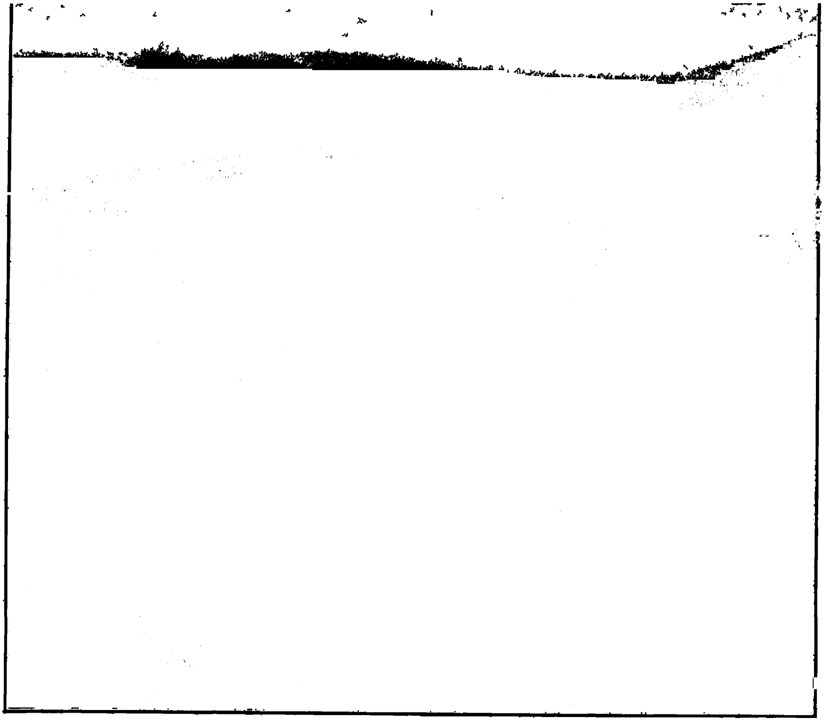


Geological Survey of Canada.

FIG. 31. Landslide topography along the Fraser River, Pavilion, British Columbia. The hummocky surface is a common feature of large slides.

in the Alps, Himalayas, and other high mountains are of frequent occurrence. In some cases man himself, by quarrying and mining operations, has unwittingly set the stage for destructive slides.

Great landslide masses that consist wholly or chiefly of loose weathered material usually move by a mechanism more complicated than simple sliding. Almost invariably the movement of such masses starts while they are saturated with water from heavy rains or from melting snow. The water greatly increases the weight of the mass and also converts all the clay and soil into stiff mud; there is localized flowage at varying rates in such a mass, although the body of material as a whole may slide on a lubricated base. Some masses of this kind move slowly and spasmodically for many years and cause little or no destruction; their motion may be thought of as excessively rapid creep.

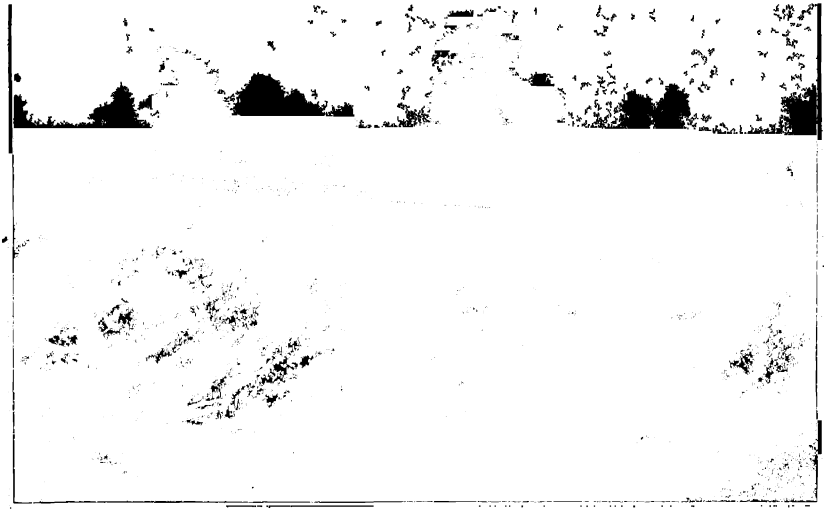


G. K. Gilbert, U. S. Geological Survey.

FIG. 32. Result of earth flowage on a steep grass-covered slope, after thorough saturation by heavy rains. Note the downslope bulge of the surface, compensating the depression made farther upslope. Berkeley Hills, California.

Some of the trees growing on one of these "chronic" landslides are likely to be tilted at various angles, and much of the surface has a characteristic hummocky topography (Fig. 31).

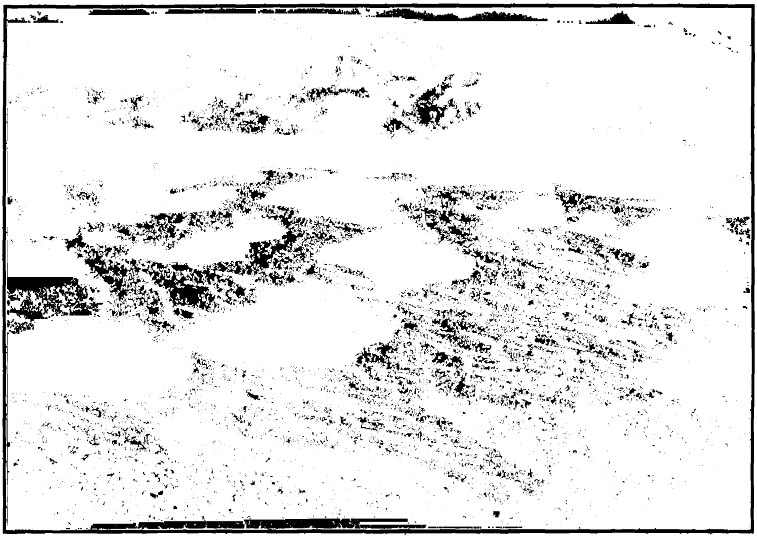
Other kinds of mass-wasting are more or less related to true landslides, although they involve also movements that resemble rapid creep Grass-covered soil on steep or moderate slopes sometimes starts flowing beneath the sod after thorough and deep saturation by exceptional rainfall Large bulges are formed downslope, and removal of support causes abrupt slumping, with breaking of the sod, on the upslope side (Fig 32) Rapid movement of soil sometimes occurs on a very gentle slope if an underlying porous layer becomes supersaturated with water Slumping of soil masses is a common occurrence along stream banks (Fig. 33), and on many steep hillsides repeated slumping produces a series of small irregular terraces; this process is in part responsible for some of the "sheep paths" that so commonly contour the slopes in hilly pasture lands.



C F. Stewart Sharpe, U. S. Soil Conservation Service.

Fig 33 Slumping of unconsolidated material in the bank of a stream, near Spartanburg, South Carolina.

Mudflows. In arid and semiarid regions, fine rock debris becomes water-soaked on steep slopes after heavy rains and moves downward as a *mudflow* (Fig. 34), which is a form of mass-wasting closely associated with running water. Whenever such a flow originates in a narrow mountain valley, the forward part continues to gather load until it becomes a stiff mixture incapable of rapid motion. This viscous frontal portion retards movement of the more liquid upstream part of the flood, and the entire mass moves along haltingly, giving out a strange roar from the grinding and bumping of boulders carried in the muddy matrix. When a mudflow of this kind passes abruptly from the confining walls of the narrow valley onto a wide piedmont slope,



Eliot Blackwelder.

FIG. 34. Marginal part of a fresh thin mudflow, after drying. East base of the Stillwater Range, Nevada.

the pent-up liquid behind the moving frontal dam breaks through with a destructive rush, and the slimy mass of muddy debris spreads into a wide sheet, ruining fertile fields and wrecking houses.

The moving mass has tremendous momentum. Houses and barns in its path are pushed from their foundations, and some large buildings have been carried as far as half a mile. Boulders weighing many tons are moved along rapidly, rolling and sliding in the muddy mixture, some of them finally coming to rest on gentle slopes far out from the mountain front.

In consistency of material and rate of movement, the mudflow represents a stage intermediate between solifluction (p. 57) and a stream of water that bears sediment in suspension (p. 78). Immense quantities of material have been moved from mountains and deposited on lowlands by mudflows in the rugged parts of Arizona, Nevada, Utah, and other western States. In southern California mudflows have been responsible for serious damage to farmlands and towns located on slopes near the mountains.

THE ROLE OF MASS-WASTING IN GENERAL EROSION

General Tendency. From the preceding discussion it is evident that the net effect of mass-wasting in all its forms is to decrease the relief of the lands. In a general way the entire mantle of weathered material on all parts of the continents is sliding or creeping steadily from higher to lower altitudes. The rate of this movement is greatest in the high, rugged mountains, least on the low, gently sloping plains; but the tendency is everywhere the same. Given sufficient time, free from renewed uplifts of plateaus and mountains, mass-wasting without the assistance of running water or other transporting agencies would eventually reduce the continents to featureless plains.

But of course mass-wasting does not go on unassisted. The several agents of erosion work together and react on each other in a most intricate way. This interaction will be appreciated better as the other aspects of erosion are explained in succeeding chapters. However, the general picture of mass-wasting can not be completed without a summary statement that anticipates parts of the later discussion.

Relation to Weathering. Most of the material moved in mass-wasting—sliderock, clay, soil, and other debris—has been produced by the weathering of bedrock. Disintegration and decomposition continue while the debris is in transit, and in many ways these weathering processes contribute to the mechanism of movement. Breaking up and softening of rock fragments in a talus remove support and precipitate sliding; continued production of clay from feldspar and other minerals makes the mantle more plastic and susceptible to flowage; the dissolving action of water not only promotes creep of the mantle but also weakens the bedrock under steep slopes and prepares the way for landsliding. Many of the movements in mass-wasting likewise increase the effectiveness of weathering. Migration of the mantle formed on hillsides exposes new bedrock to mechanical and chemical attack. Rock fragments moved in mudflows or landslides are broken by impact and grinding, and thus fresh surfaces are exposed. When great masses of bedrock are moved and broken up in landsliding, as at Frank, Alberta, enormous quantities of new material are made available to active weathering.

Relation to Running Water. Part of the water from every rain flows on the surface, either in thin local sheets or in definite channels (p. 73). This water carries debris directly and also promotes in many indirect ways the progress of mass-wasting. Washing of weathered material from a talus disturbs equilibrium and sets the sliderock in motion. When gullies are cut many of the steep banks slump, and creep is promoted in the adjacent mantle. The rugged topography of highlands results in large part from the cutting of deep valleys by streams (supplemented locally by glacial erosion); and the most active mass-wasting, by sliding and creep, takes place on the sides of mountain valleys.

The relationship is reciprocal, for mass-wasting is a potent auxiliary of running water in the work of leveling the lands. Most of the rock waste contributed by the continents to the sea floors is delivered ultimately by streams (Chap. 5). However, a large part of this waste is brought within reach of streams by creep and other forms of mass-wasting (Fig. 29). In highlands, landslides as well as rapid creep deliver material directly to large streams and their tributaries, locally more rapidly than it can be removed. In lowlands the alluvial banks slump into or toward the streams (Fig. 33), and creep operates over a wide area even though its rate is slow. In all kinds of topography, removal by streams of the creeping mantle as it reaches the valleys prevents clogging of the creep mechanism and keeps the system of supply in operation. Thus running water and mass-wasting are closely geared in the scheme of erosion.

Mass-wasting is further explained and illustrated in connection with processes discussed in later chapters.

EFFECTS OF WEATHERING AND MASS-WASTING ON LANDSCAPES

Major Effects. In a journey from the eastern United States to the arid Southwest the traveler is impressed with the striking contrasts in the landscapes. These differences reflect complex causes, but among the most important factors are rate and kind of weathering and processes of mass-wasting, under the different climatic conditions. The abundant vegetation of the Eastern States in itself creates a strong contrast with the barrenness of New Mexico and Arizona; but there is a more fundamental difference in the actual land forms. In the East the hills and slopes are characterized by rounded curves and smoothed profiles; in the Southwest the profiles are angular.

Disintegration as the dominant type of weathering tends to produce angularity. Forces that break the rocks take advantage of original joints which form square-cornered blocks, and, as these natural units loosen and fall, the cliffs from which they are displaced stay rugged and steep. Ridges and hills made of flat-lying strata remain flat topped and steep sided as they grow smaller through disintegration and recession of the cliffs. Extensive flat-topped erosion remnants of this kind are called *mesas;* similar features of smaller extent are the *buttes* of western landscapes (Fig. 35). Wherever the structure of the rocks favors development of sharp peaks and jagged ridges, these features tend to keep their irregular forms as they waste away. Divides between adjacent valleys are either flat topped (where strata are horizontal), or sharp and rugged, and valley sides are steep and commonly step-like, with a succession of cliffs (Fig. 11). These are some of the typical forms of the arid and semiarid southwestern United States, where there is little decomposition and soils form scantily.

Vigorous chemical weathering rounds off sharp corners in both large and small features of the landscape. Vertical cliffs and angular ridges in a humid region are the exception rather than the rule; their presence

SPENCE AIR PHOTOS.

Fig. 35. View in Monument Valley, Utah. Sandstone nearly 400 feet thick, once continuous over the area, has been eroded; remnants form a large mesa (at right) and isolated buttes. Weak shale beneath the sandstone underlies the gentler slopes at the base of Mitten Butte (middle background). Vertical fractures in the sandstone outline slender, chimney-like remnants.

I. C. RUSSELL. U. S. GEOLOGICAL SURVEY.

Fig. 36. Rounded hilltops and smooth slopes characteristic of a region with a moist climate. Deep soil mantles most of the surface. The small terrace-like features in left foreground are paths made by generations of grazing animals. North of Luray, Virginia. Note contrast between this landscape and the one shown in Fig. 35, which is characteristic of a semiarid climate.

indicates either brief exposure to the atmosphere or extremely resistant rocks. Sharp peaks and serrate ridges tend to disappear under the influence of decomposition. Mantle forms on valley sides, and as it accumulates and slowly migrates from higher to lower levels the slopes take on smooth curving profiles (Fig. 36). With the formation of soil and the spreading of vegetation, decomposition becomes more effective and the slopes grow still smoother, because the cover of vegetation protects the slopes from gullying by running water, and there is built up a nearly universal layer of soil even on the steep slopes. On opposite sides of a ridge cliffs retreat, grow lower, and disappear, and the top of the ridge becomes smooth and rounded under a cover of soil. Thus divides between adjacent valleys tend to become broad and smooth in contrast to the angular divides of arid regions.

Farming Problems. Great quantities of rich soil are lost every year through the combined operations of mass-wasting, running water, and wind. Soil conservationists recognize the seriousness of the problem; but they find it much simpler to check the destructive effects of slope wash and gullying than to control the large masses of earth usually involved in creep and slumping.

There is hope that some of the most immediate problems can be at least partially solved by combination of direct and indirect methods. For example, in semiarid parts of some western States overgrazing of range lands has increased the prevalence and destructiveness of mudflows, which not only remove the topsoil from large areas but, after mixing this soil with unproductive clay and gravel, bury the rich topsoil of lower areas under the infertile mixture, thus ruining all the ground affected by the movement. The only effective remedy is to restore the protective sod, which requires a comprehensive conservation program to prevent the recurrence of overgrazing. Slumping of farm land adjacent to deep gullies can be checked by measures, now extensively practiced, leading to gradual filling of the old gullies and prevention of further gully development. Slumping and active creep in some rich bottom lands can be stopped by protecting adjacent stream banks from erosion (Fig. 33).

Engineering Problems. Knowledge of the principles governing mass-wasting is essential in the selection of successful sites for large buildings, dams, and other engineering works. Durham Castle has stood remarkably well, in view of its location (p. 56); numerous structures built more recently have collapsed as the result of creep and other movements of the mantle. The St. Francis dam in southern California, which failed with appalling loss of life and property (p. 79), had one of its abutments in ground subject to sliding—a fact that should have been recognized at the time the damsite was selected.

Some engineering works have to be constructed in spite of difficulties from mass-wasting. When the Culebra Cut of the Panama Canal was under construction, the deep artificial excavation caused sliding, slumping, and flowage of the adjacent ground on such a large scale that for a time the project appeared to face defeat. It was necessary to take out an enormous yardage in excess of the original estimate; and after the canal was open for use there was recurrent trouble in the Culebra section. Some landslide masses can be stabilized by construction of drainage tunnels to remove water, which plays an essential part in the movement. In constructing the Grand Coulee dam, on the Columbia River, a slide in wet silt was started by excavation and gave considerable trouble until engineers hit on the ingenious plan of artificially freezing the base of the slide and thus stabilizing the ground until they could complete the adjacent part of the foundation.

Private companies and governmental bureaus are put to enormous yearly expense in repairing and rebuilding railroads, highways, aqueducts, and other communication lines that are injured by landslides and similar movements in rugged country.

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Describes a number of typical rock glaciers in Alaska and analyzes the mechanics of their growth and movement.

CHAPTER 5

RUNNING WATER

GEOLOGIC IMPORTANCE OF STREAMS

Stand on the brink of the Grand Canyon of the Colorado (Fig. 11), and look down at the winding river a mile below. The rock-walled canyon is impressively vast, and by contrast the stream at its bottom seems puny. Those who first speculated on the history of the canyon thought that the Colorado had found this low-level course already prepared for it by a great gash or rift earlier opened by some cataclysm in the Earth's crust. But today we are convinced that the river once flowed through a shallow trough at the level of the canyon's brink. We can see that very gradually the sediment-laden water has cut down through the underlying rocks, sawing ever more deeply until in the course of time, and with the aid of mass-wasting, it has dug a trench a mile deep and several miles wide. Looking at the intricate pattern of canyons and subcanyons which, arranged like the veins in a leaf, join the main valley, most of them entering at grade with the main stream, we realize that they were developed not by chance but as members of a unified and interrelated system, obeying a common law and having a common destiny. And when we see this same pattern repeated again and again in stream systems large and small throughout the world, we conclude that all streams act according to certain definite laws imposed on them by their surroundings.

Streams return to the sea the water evaporated from it, carried by the atmosphere over the lands, and there precipitated as rain and snow. Furthermore streams as a group, coupled with the mass-wasting processes with which they are integrated, constitute the most important single agent of erosion on the face of the Earth. Continuously excavating valleys and continuously carrying away the excavated rock waste, plus the far greater contributions from mass-wasting of the land, streams will continue to operate as long as there are lands with rain falling upon them. The processes of excavation and washing away, although complicated, are based on a series of principles which enable us not only to follow intelligently the intricate details of sculpture of the land but even to predict to some extent the way in which the land will be carved in the future.

STREAMS AND STREAM FLOW

Precipitation. Probably there is no part of the Earth's surface on which rain or snow does not sometimes fall. Even the most arid spots in the world's driest regions receive a little rain once every few years. Death Valley in southeastern California, one of the driest areas in North America, has about 2 inches of rainfall annually. Whereas the thickly populated temperate regions such as eastern North America and western Europe have 20 to 60 inches of rain (Fig. 17), some parts of India receive as much as 500 inches every year. In some places the time distribution of rainfall is more significant than its annual amount. Thus if total rainfall is concentrated in a few intense storms, it causes a series of floods, and the work done by the water in eroding the land is far greater than it would be if the same amount of rain were distributed so nearly uniformly throughout the year that floods did not occur.

The variable distribution of precipitation both areally and in time, therefore, exerts a strong influence on the sculpture of the land and accounts in a large measure for the differences of landscape in various parts of the world.

Runoff. A part of the precipitation that reaches the surface of the land evaporates and returns to the atmosphere, a part sinks below the surface and becomes *subsurface water* (Chap 7), and a part flows down the slopes of the surface, forming the *surface runoff*. Much of the subsurface water later emerges as *ground-water runoff*, adding greatly to the runoff as a whole.

About one-third of all the precipitation that falls forms runoff directly or indirectly. At any one place, however, the proportion contributed to runoff depends on several factors such as (1) surface slope, (2) permeability of the local rocks, (3) character and amount of vegetation, (4) temperature and humidity of the atmosphere, and (5) amount and distribution of the precipitation throughout the year. Therefore runoff varies greatly from place to place Some areas of sand and very permeable limestones have essentially no surface runoff, because all the precipitation sinks below the surface. On the other hand, in a country such as the Appalachian region nearly all the water from rains in the early spring runs off because the ground is already saturated with melted snow.

Streams. No land area is either perfectly flat or perfectly smooth. As a result the surface runoff, though it may start as *rainwash*—a thin sheet of water flowing evenly downslope—is quickly concentrated by converging slopes into the shortest and steepest routes downward. Thus, gradually augmented by ground-water runoff in the form of seeps and springs, the runoff becomes organized into definite *streams*, which might be defined as water and rock waste streaming toward the sea along more or less definite courses. Each stream occupies a valley, excavated or modified by itself, of which it is an integral part. Most valleys are floored with rock waste deposited by their streams, and usually these partial fillings are saturated with water, which percolates down-valley at rates much slower than the flow of the surface waters which they support.

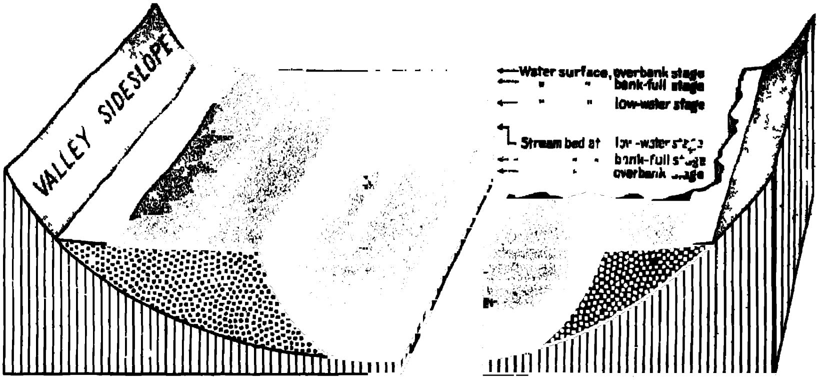


FIG. 37. Segment of an ideal valley, cut to show three standard positions of the stream surface and three corresponding positions of the stream bed. Natural levees are not shown. (Vertical scale greatly exaggerated.)

We have said that about one-third of the world's precipitation becomes runoff. Much of this water streams seaward from altitudes of thousands of feet. The energy that results is tremendous; the work it performs is very great. Let us examine the manner in which this work is done.

Discharge. Every stream occupies a *channel*, a water-filled groove which in a narrow valley may include the entire valley floor but which ordinarily occupies only a small fraction of the valley floor (Figs. 37, 42). The flow of water through the channel is expressed as *discharge* (the volume that passes through a given cross-section of the channel during a unit of time), stated in cubic feet per second. Discharge measures the amount of water fed to the stream from surface-water and ground-water runoff.

The casual observer is likely to judge the "might" of a stream by its width, the only dimension usually visible to him at any one place. He is often misled, because some narrow streams, by virtue of steep gradients or deep channels, discharge much more water than do some wide streams. These differences are controlled by two principal factors: (1) *velocity of flow,* and (2) *cross-sectional area of the channel.* Velocity, in turn, depends on (*a*) *gradient* (downstream slope, usually expressed in feet per mile), (*b*) *volume of water,* (*c*) *load* of rock particles being carried, and (*d*) *shape of the channel.* This last factor is important because friction between water and channel retards stream flow. Hence greater velocity is possible (other factors equal) in a channel having a semicircular-shaped cross-section than in a broad shallow channel with the same cross-sectional *area,* because the former shape offers, in proportion to its area, the smallest possible surface on which friction can occur.

Discharge equals velocity times cross-sectional area. Velocity varies approximately as the square root of the gradient. Although velocity commonly increases also with increased volume, the relationship can not be expressed as a simple formula.

Because of the interrelationships mentioned, these various factors are adjusted to each other by the stream itself, which alters its channel by erosion and deposition from place to place. For example, in most streams discharge increases from head to mouth, reflecting increments from tributaries and from ground-water runoff, whereas the gradient decreases from head to mouth. The apparent paradox is explained chiefly by (1) the variable cross-section of the channel, which increases downstream sufficiently to permit greater discharge despite reduced velocity of flow, and (2) the shape of the channel, which commonly becomes deeper relative to its width as it is followed downstream.

An unusual example is the Clark Fork of the Columbia River in northeastern Washington. At the town of Usk this stream is 2000 feet wide and 15 feet deep, with a gradient of less than 1 foot per mile. The mean discharge here is about 15,000 cubic feet per second. At Z-Canyon, 45 miles downstream, the same river is 18 feet wide and 175 feet deep, with a gradient of 60 feet per mile. The discharge is about 15,500 cubic feet per second. Although the Clark Fork looks much smaller at Z-Canyon than at Usk, actually its discharge is slightly greater. More water passes through Z-Canyon, despite the smaller channel capacity there, mainly because the gradient is much steeper.

Within a stream channel, velocity of flow is greatest where friction is least. Hence velocity is greater along the center than along the sides and is greater also near the top than at the bottom of the channel.

Long Profile. The continuous curve formed by the slope of a stream from source to mouth is its *long profile.* Although the long profile of a newly formed stream is the profile of the land surface itself, the long profile of a well-established normal stream is a curve, concave up (Fig. 38), that tends to approach a hyperbola. This results from adjustment of each part of the profile to the conditions in the corresponding part of the stream. In its upper course the stream's discharge is slight, but, owing to the usually steep initial slope of the land, its gradient is steep. It is not fully loaded; hence most of its energy is applied to downcutting. In its lower course discharge is much greater; for this reason and because its channel is more efficient and its load is finer grained, it can move its load on a gentler gradient, and downcutting is at a minimum. Intermediate conditions obtain at intermediate points; hence the gentle slope downstream gradually becomes a steep slope upstream. With the passage of time, as the stream cuts down toward baselevel (p. 81), the long profile becomes gradually less steep throughout its length.

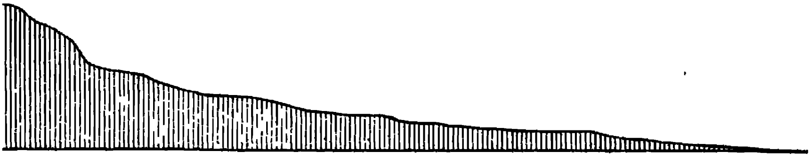


FIG. 38. Long profile of the Lehigh River and its continuation through the lower Delaware River to the sea, illustrating the general concave-up form of its curve. Baseline = sealevel. Long dimension, 120 miles. Altitude of stream at top of profile, 820 feet. Vertical exaggeration, 132 times. The irregularities in the profile are caused by unequal resistance of the rocks to erosion and possibly by other factors.

Floods. Not only does a stream adjust its channel from place to place; it also alters its channel from time to time in the same place as a result of variable discharge, which usually varies with the seasons. Most streams are subject to flood at times of most concentrated rainfall or greatest thawing of winter snow. Floods are measured at selected points by gages which record the height to which the water rises. The approximate cross-sectional area of the channel being known at these points, the discharge can be calculated. In this way it has been determined that in the Mississippi River immediately below the mouth of the Ohio, the normal rise of the water surface in spring floods is about 40 feet, while the corresponding flood discharge is between 10 and 15 times the normal low-water discharge. On small streams, differences may be even greater. For instance, during a recent flood the discharge of Tiger River near Woodruff, South Carolina, increased from normal to 24 times normal within 4 hours. Twenty-four hours later it decreased again almost to normal.

A stream flood is essentially an upward bulge of the stream's surface, analogous to the bulges in the surface of the sea caused by the tides (p. 224). This is clearly shown by the gages, which rise to their maxima one by one as the crest of a bulge passes downstream. Many stream channels are capacious enough to carry the discharges of ordinary floods. When the water rises to the tops of the stream's banks, filling the channel completely, the stream is said to be in *bank-full stage* (Fig. 37). Further rise causes the flood to overtop the channel and spread out on the adjacent valley floor; it is then said to be in *over-bank* stage. The spread-out water flows so slowly, owing to the great friction on its base, that it often appears ponded; yet the velocity in the submerged channel is usually great.

STREAM EROSION AND DEPOSITION

Hydraulic Action. Hydraulic action, abrasion, solution, and transport together constitute stream erosion, and deposition is closely related to all of them.

The force inherent in the flow of water is able to lift and move away loose material. This is illustrated by the impact of water from a garden hose playing upon loose soil. The soil is churned up and washed away, leaving tiny gullies excavated by the escaping water, and for the entire operation hydraulic action is almost solely responsible. Every rainfall accomplishes the same result in plowed fields and on steep slopes, and in all streams hydraulic action operates as an important factor, although the effects attributable exclusively to it are usually masked by the effects of other processes that are going on at the same time.

Hydraulic action is an important factor in quarrying out blocks of rock along cracks and planes of weakness. This type of erosion is clearly apparent in a stream bed consisting of thin horizontal layers of rock. The force of the flowing water, no doubt aided largely by frost wedging, loosens blocks which are then quarried out and carried away by the current. In places where this process is dominant, the layers of rock are stripped back successively to form flights of broad low steps descending downstream. Where the rocks are cut by closely spaced vertical joints, as is true of many lava flows (p. 372), a stream may quarry out joint blocks unevenly, producing an extraordinarily rough, irregular surface.

Abrasion. Abrasion is the mechanical wear of rock on rock. In streams it is caused by friction between rock fragments moving with the stream and rock fragments or firm rock in the stream bed, and by friction between fragments moving through the stream at different rates. Rubbing, knocking, bumping, scraping, grinding, scouring—all these terms are applicable to stream abrasion. This process depends on the presence of rock fragments in the stream, whereas hydraulic action can be accomplished by clear water. On the other hand, no stream is entirely free from rock fragments, for it acquires them through its own hydraulic action, through contributions from tributaries, and through slumping of its banks. Both hydraulic action and abrasion operate in all streams and work in close co-operation with each other. Not only does abrasion wear down bedrock exposed in the stream; it also wears down the rock fragments themselves, making them smaller and rounding them more and more (p. 256).

Abrasion is well illustrated by bowl-shaped or cylindrical *potholes*, ranging from a few inches to many feet in diameter and depth, in the rock bed of a swift stream. Most potholes are excavated by rock fragments swirled about by persistent eddies. Abrasion of the grinding tools is shown by the rounded shapes of the pebbles commonly found in the potholes. The valleys of some small and very steep streams have been excavated largely by pothole abrasion.

Solution. All streams contain substances in solution, acquired in part from subsurface water and in part directly from surface runoff. Limestone (calcium carbonate) is very susceptible to the solvent action of stream water. But far less soluble rocks are affected. The water slowly decomposes some of the minerals and thereby loosens the constituents of the rock and prepares them for mechanical seizure by the current. However, only a small proportion of the matter dissolved in stream water was dissolved by the streams themselves. By far the greater part of it was dissolved by subsurface water and was contributed to the streams later in the ground-water runoff.

Transport and Deposition. The rock waste carried by a stream and the substances dissolved in the water together constitute the stream's *load*. The load is transported by the hydraulic action inherent in stream flow. Flow, however, is not uniform throughout the channel but is turbulent, involving many cross currents and eddies, and the turbulence increases rapidly with increasing velocity. Consequently loose rock particles on the bottom are lifted and kept in motion by repeated tossing into the body of the stream as they are carried forward. The smaller and lighter particles are tossed up easily and frequently and so travel steadily downstream. The larger and heavier fragments are rolled or pushed along the stream bed, but movement of this kind is necessarily much slower and less continuous than that of suspended particles. If the velocity of the stream decreases slightly, the largest and heaviest of the fragments dragged along the bottom are no longer moved at all; accordingly they are left behind and accumulate, while the smaller particles move on. For this reason nearly all stream deposits are sorted according to size and specific gravity (p. 264). Only rarely is the velocity of a stream checked so suddenly that particles of markedly different sizes are deposited heterogeneously together; this sometimes happens where a stream emerges from a mountain district on to a flat plain and builds a fan (p. 87).

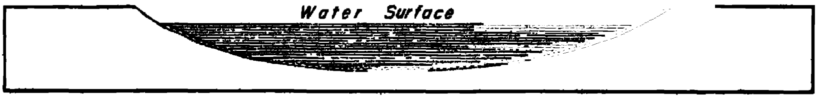


FIG. 39. Diagrammatic cross-section of stream channel showing its *wetted perimeter* (beaded line) and *cross-sectional area* (shaded).

Relation of Velocity to Erosion and Transport. Stream velocity increases with increasing gradient. It increases also with increasing discharge, as during the floods to which most streams are subject. The increase can not be stated in exact terms because the presence of variable factors makes an accurate formula impossible. However, empirical formulas, which approximate the relationships, have been devised. One of the best known states that $V = B\sqrt{rs}$, in which V = velocity, B = a constant, the value of which varies from stream to stream, s = the slope or gradient, and r = the ratio of the cross-sectional area of the channel to the length of the wetted perimeter of the channel (Fig. 39).

An increase in velocity increases the ability of a stream to erode its bed and transport a load. In theory, under ideal conditions, doubling the velocity may (1) increase abrasive power about 4 times; (2) increase the diameter of the largest piece of rock the stream can push along its bed by as much as 4 times; and (3) greatly increase the capacity to transport rock fragments of a given size.

These powers result from doubling the velocity, through increase either of gradient or of discharge. If the gradient is increased 4 times, velocity is about doubled. But greater-than-fourfold gradient increase is common between the broad tops of many hills and their steep sides. When hills of this kind are washed by the immediate runoff from rains, serious soil erosion is likely to occur if the slopes are bare and unprotected (Fig. 49). Again, the increased discharge of a stream in flood often much more than doubles its velocity. During normal seasonal floods, the velocity of the Mississippi at points along its lower course increases 2 to 4 times. Increases of velocity equal to 10 to 20 times the mean velocity are not uncommon under flood conditions. For example, at Montague, Massachusetts, the velocity of the Connecticut River, a major stream, was increased by 21 times in a single flood. Such huge increases of velocity, resulting from increases of discharge alone, make it clear why most streams perform the great bulk of their eroding and transporting work during flood stages, and most of their deposition of rock waste as the floods subside.

The effect of increased discharge upon volume of load is illustrated by the San Gabriel River at Azusa, California. During a flood in 1920 the discharge of this stream increased 20 times, while its silt load shot up from little more than zero to 55 per cent (by weight) of the water. The effect of increased velocity on the ability of a stream to move large pieces of rock is shown by the case of the St. Francis dam near Los Angeles. In March, 1928, this dam gave way and sent a sudden flood rushing down the valley below. This flood moved blocks of concrete weighing 10,000 tons each.

In India, during the Gohna flood of 1895, which lasted just 4 hours, the water picked up and transported such quantities of gravel that through the first 13 miles of its course the subsiding stream made a continuous gravel deposit 50 to 234 feet thick. In each of the instances cited, such work was impossible under low-water conditions.

Adjustment of Channel and Load to Discharge. It has been stated that gradient, and size and shape of the channel, are continuously adjusted to fluctuations in discharge. The adjustments are made by means of erosion and deposition and are therefore reflected by changes in the load. Increasing discharge increases the velocity, and hence the abrasive power of a stream, as well as the diameters and total volume of the rock fragments it can transport. As a result, the channel is scoured out, particularly near its center line where velocity is greatest, and is thereby deepened relative to its width and thus made more efficient. The rise of water recorded on the flood gage therefore records only a part of the increased cross-sectional area of the channel. The rest of the increase is expressed by the corresponding deepening of the channel, which takes place at the same time, and which may amount to two or three times the rise. The material excavated is transported downstream as part of the load, and, as the flood subsides, this is dropped on the bottom in the order of size and weight, the finer and lighter fragments settling last.

Thus a stream in flood scours out (*degrades*) its channel and partly refills (*aggrades*) it with rock waste as the flood subsides. In some large streams, scour and fill remove and then replace sediment to a thickness of 100 feet or more with each flood (Fig. 37).

At any stage, rock fragments are continually being picked up by the turbulent stream and others are dropping out. In a given stream the number and the sizes of the fragments in motion at any moment are limited by the stream's energy. As discharge increases, velocity and energy increase, more and larger pieces are picked up, and fewer are dropped out. This causes erosion of the channel floor, which is thereby lowered. But the channel floor is not lowered at its mouth, because the gradient at that place is zero. Hence lowering of the channel at any point above the stream's mouth must reduce the gradient between that point and the mouth. This reduction of gradient tends to reduce the velocity and to limit the increase in velocity brought about by the increased discharge.

When this limit is reached no more erosion occurs, and the equilibrium between the pieces picked up and those dropped out is maintained until the discharge begins to diminish, at which time excess of deposition over erosion aggrades the channel floor. This increases the average gradient to a degree sufficient to enable the stream to transport its remaining load without further upbuilding.

Thus by erosion or deposition, the stream continually adjusts its gradient so that it can just transport the load to be carried. Such adjustments occur with every flood, but they occur also with major changes in the condition of the stream. For example, the Colorado River carries a bulky silt load, for the transportation of which a steep gradient is necessary. The completion of Hoover Dam in 1935 caused deposition of this load in the lake above the dam; the river below the dam, robbed of its load, required a gentler gradient than before, and accordingly it reduced its gradient there by conspicuous erosion.

The system of flood- and low-water stages of a stream—fluctuations (usually seasonal) through a norm—with the delicate channel adjustments that accompany these systematic changes, is the stream's *regimen*. In general terms the regimen of a stream is its behavior or habit throughout the year. Ordinarily a regimen is overthrown, and a new one substituted for it, only by major external factors such as movement of the Earth's crust beneath the stream (p. 355), a volcanic eruption (p. 304), or glaciation of the stream's valley (p. 188).

Baselevel. A stream can continue to erode its bed until it reaches the sea. Here its current is checked, preventing further downcutting. The sealevel, therefore, is a level below which, in general, streams can not cut (although some powerful streams like the Mississippi are able to keep parts of their channels scoured out to somewhat greater depths). The level of the sea, projected inland as an imaginary plane below the surface of the land, is termed *baselevel,* because it is the base that limits downward erosion by streams.

If a stream empties into a lake, its downcutting is limited by the level of the lake for as long a time as the lake exists. The lake therefore acts as a local baselevel for the stream, but because every lake above sealevel must be drained ultimately, the baselevel that it represents is temporary as well as local. For these reasons the level of the lake, projected inland, may be regarded as a *local and temporary baselevel* controlling the region upstream from it. Similarly the floors of lakeless basins in arid regions (p. 115) act as local and temporary baselevels for the streams tributary to them. Barriers of resistant rock across the path of a stream limit in the same way the extent of downcutting upstream from them. All local and temporary baselevels, however, disappear eventually, and the areas they formerly affected pass into the control of the ultimate baselevel.

Profile of Equilibrium; The Graded Stream. We have seen that there is a mutual adjustment between erosion and deposition, and the load in transport. In the early stages of valley excavation and along the upper courses of most streams, downcutting is the dominant process. As downcutting reduces the gradient and the products of downcutting are added to the load, a time comes when the increasing burden of transporting the load requires energy that was formerly applied to downcutting. Downcutting is retarded just to the extent necessary to enable the stream to transport its load. Thus the stream adjusts its gradient to its load and tends to approach a condition of balance between the available energy and the work done in transporting the load. In other words the long profile approaches a *profile of equilibrium.* However, frequent variations in discharge and other factors prevent the profile of equilibrium from ever being actually attained. Yet, because the tendency toward its attainment is ever present, the stream is able gradually to eliminate falls, rapids, and other irregularities and thereby to make its long profile smooth. When a near-equilibrium profile of this kind has been evolved, the stream is said to be *graded.* Because near-equilibrium conditions of energy and load may be attained in some parts of a long profile sooner than in other parts, the graded condition is reached sooner in some places than in others.

Nearly all graded streams have beds consisting of loose sediment. Streams that flow on bedrock are rarely graded because their loads are small and much of their energy can be devoted to erosion of the bedrock. Unless a change in the regimen occurs, a stream maintains its graded condition indefinitely, while very slowly reducing its gradient.

As a rule a tributary enters its main stream at the level of the main, in spite of the greater downcutting ability of the main caused by its greater discharge. Most tributaries are able to maintain this accordant relationship because every foot of downcutting by a main stream increases the gradients of its tributaries and permits the tributaries to "keep pace" in spite of their smaller discharges.

The increased gradients of the tributaries augment downcutting at the expense of lateral cutting, and this may result in very deep narrow tributary valleys. In their effort to maintain accordant relations with the rapidly downcutting Colorado River, some of its tributaries have cut slot-like canyons with vertical sidewalls.

When a part of a main stream reaches grade, the local tributaries soon become graded with respect to it.

Any long-term change in gradient, discharge, or load (exclusive of seasonal changes within the regimen) upsets the graded condition by altering the rate of erosion. Graded streams have been converted into actively aggrading streams through great increases in load caused by overcultivation of farm land, overgrazing, or deforestation, or when glaciers, appearing in their headwater regions, poured great additional quantities of rock waste into them.

Slower changes of longer duration occur when a movement of the Earth's crust warps a land area drained by streams and so alters their profiles (p. 485). The streams begin at once to adjust themselves to the new profiles and sooner or later again reach a graded condition that is also adjusted to the altered position of the land. All streams therefore work toward the near-equilibrium implied by the graded condition; if thrown out of it, they slowly but surely return to it; and, if left unmolested by interruptions, they would maintain it as long as they continued to flow.

Falls and Rapids. Falls and rapids occur at points where gradients are abruptly steepened. There is no sharp distinction between a rapid and a falls; both commonly occur where streams encounter previously existing cliffs or pass from resistant rocks to weak rocks. Niagara Falls, for example, was formed by a river that fell over a ready-made cliff (p. 194). The Niagara River, originating as the outlet of Lake Erie at Buffalo, flows over a plateau that ends near Lake Ontario in an escarpment (Fig. 40). At the surface of the plateau are layers of resistant dolomite, and underneath are weak, easily eroded shales (Fig. 41). When the river came into existence by the overflow of Lake Erie, it flowed across the plateau and tumbled over the escarpment at Lewiston, forming a falls. As the falling water gradually undermined the weak shale, the resistant dolomite above was left projecting as a lip, which, penetrated by cracks and fissures and left unsupported, fell away block by block. Through long continuation of this process the falls perpetuates itself and gradually retreats upstream, leaving a great gorge downstream to mark its path. It has now moved 7 miles back from its original position and is still visibly retreating. On January 17, 1931, a large mass of rock at the lip of the American Falls was undermined and fell, leaving the lip greatly altered. Similar masses

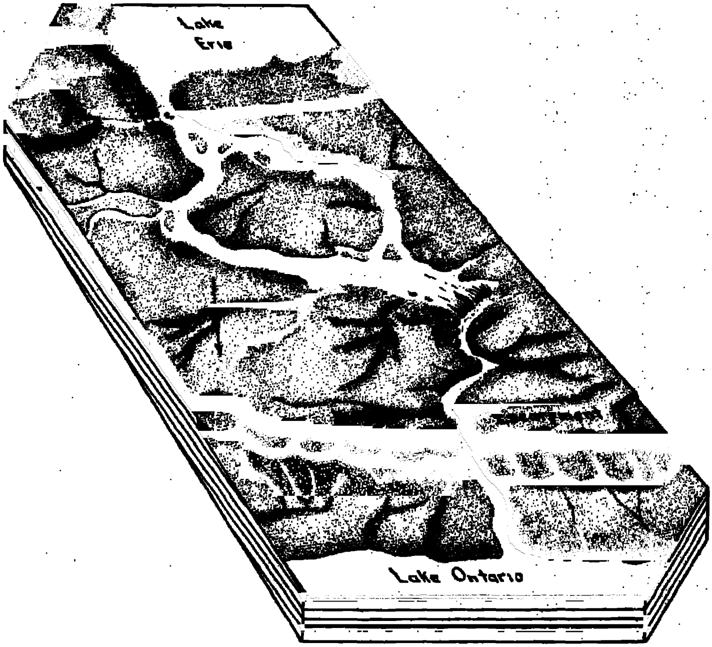


FIG. 40. Bird's-eye view of Niagara River and Falls, looking south. Greatest length of block, 35 miles. (Vertical exaggeration, 2 times.)

weighing thousands of tons fell from Horseshoe Falls on August 13 and December 5, 1934, and in September, 1938. During the 85-year period 1842–1927 Horseshoe Falls retreated at an average rate of 3.4 feet per year.

Not all falls, however, were formed by ready-made escarpments. More rapid erosion in weak rock than in resistant rock lying immediately upstream will result, at the contact, in a steepening of the gradient that may become a falls.

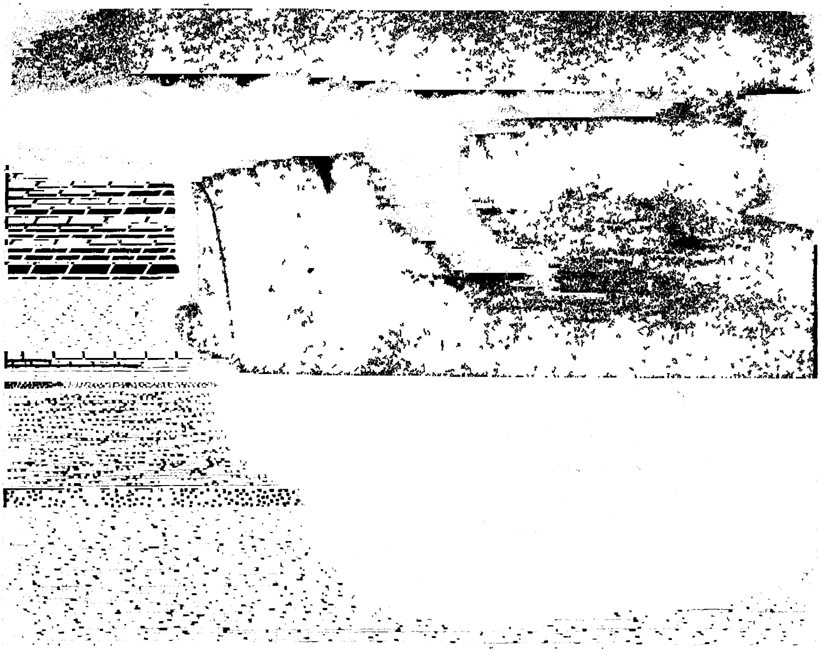


FIG. 41. Relation of Niagara River to bedrock at the Falls. Section shown is on the northeast edge of the American Falls. Goat Island in middle distance; Horseshoe Falls in background. Conditions below surface of Niagara River are partly inferred; probably the gorge was excavated chiefly by the Horseshoe Falls rather than by the smaller American Falls. The overdeepened part of the valley is a *plunge basin*, excavated not by the comparatively weak American Falls but by the more powerful Horseshoe Falls before it had receded to its present position.

All falls are temporary. They are eventually eliminated by stream erosion as stream profiles are reduced to the graded condition.

It is worth repeating that falls and rapids occur only where streams are not graded. When a stream becomes graded, sharp irregularities in its long profile, such as falls and rapids, are necessarily smoothed out and destroyed.

STREAM DEPOSITS

TYPES OF DEPOSITS

Stream deposits (or, as they are often called, *alluvial deposits* or *alluvium*) take many different forms, even in a single valley, and constitute in themselves an extensive subject of study pursued by engineers and geologists concerned with river navigation, flood control, and dams for making electric power. Without going into details, we can only indicate the common types of stream deposits and the general manner in which each is made.

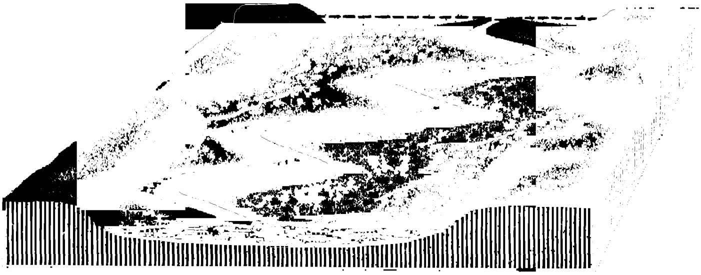
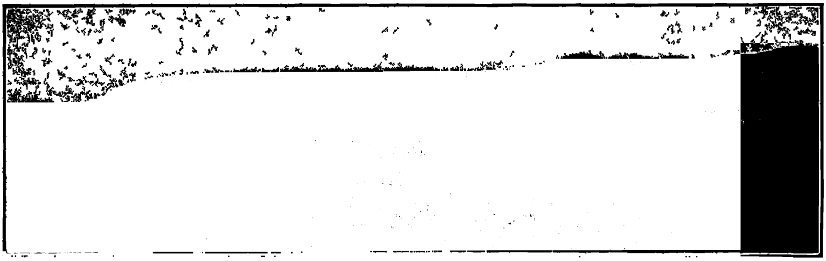


FIG. 42. Valley floor with natural levees bordered by swamps. The vertical scale is exaggerated, and the stream and levees are shown larger in proportion to the width of the valley than they would normally be. Dashed line shows position of water surface in time of highest overbank floods.

Channel Deposits. When a stream *meanders* (p. 102), the current is deflected against first one bank and then the other. Deflection of the current to the outer and downstream sides of the channel creates slack water at the inner and upstream sides and causes deposition there which builds those sides out, tending to narrow the channel and causing the current to cut the opposing sides away (Figs. 56–58). This results in migration of the meander and in the spreading of a veneer of deposits, which in time covers the entire valley floor. Although they may be very extensive, such deposits ordinarily are no thicker than the stream is deep. They are made strictly *within the channel.*

Flood Deposits. A stream whose floods reach the overbank stage rises and inundates the adjacent valley floor. When the overflow is forced out on to the flat, its velocity is quickly and effectively checked, causing concentrated deposition of fine-grained rock waste along the immediate borders of the flooded channel, gradually growing less in an outward direction away from the channel. The deposits therefore take the form of low but distinct ridges (*natural levees,* Figs. 42, 43) that border the channel and remain after the flood has subsided. The ridges have heights from 4 feet up to 15 feet or more; and, although they are the most pronounced visible features of some valley floors, their heights are small compared with the depths of the channels. In the course of such flood deposition the valley floor is capped by a thin mantle of fine sediment deposited quickly in somewhat lake-like shallow flood water.

The part of a valley floor that is subject to overbank floods is termed a *floodplain*. The Mississippi River has a floodplain which, including the natural levees and the low swampy lands beyond their outer slopes, formerly covered an area of 35,000 square miles, but has been reduced by artificial means to about 25,000 square miles.



Geological Survey of Canada.

FIG. 43. Valley of the Kootenay River, British Columbia, during a great flood. The valley floor is entirely under water except for the trees growing on natural levees along the main channel.

Many stream valleys, such as the lower reaches of the Connecticut River valley in Massachusetts and northern Connecticut, have both meander deposits and floodplain deposits.

Short Cuts; Sloughs. Where natural levees are inconspicuous or absent, flood water overflowing a meandering channel follows some old abandoned channel or excavates a new channel across the tongue of land formed by the meander. Some of these short cuts become enlarged, permanently cutting off the meanders they span. Most, however, are abandoned as the flood subsides and are left as *sloughs,* which are slowly undercut as the meander shifts downstream.

Deposits of Braiding Streams. During the low-water stage, many streams heavily loaded with fine sediment, or losing volume by evaporation or by percolation into their beds, choke up their channels with deposits; they then overflow and cut new channels, which in turn become rapidly choked. This process, termed *braiding,* develops an intricate network of shallow channels forming a complex pattern on the valley floor. Many streams in arid and semiarid regions (for example, the

SPENCE AIR PHOTOS.

Fig. 44. Fans being built out into Death Valley, California, from Black Mountains (left). View south from Furnace Creek Ranch (right foreground), showing Death Valley, a closed basin with a white salt-incrusted playa, and the Panamint Range inclosing the basin on the west (right). The relations shown here are essentially those portrayed in Fig. 75.

South Platte River in western Nebraska) and streams of glacial meltwater heavily loaded with sediment are intricately braided (Fig. 124) The fact that a stream fills its channel with deposits at low water doe not imply that the valley floor as a whole is being built up, becaus during flood stages the stream may remove more rock waste than i deposits during low water.

Bars. Bars of sand and gravel are built in the channels of man streams in which the load is not great enough to cause braiding. I some streams, bars form during low water and are swept away by th next flood. In others they are continuously present, shifting their for and position from season to season.

Fans. When a swift tributary stream flows down from high lan on to a wide and nearly level valley floor, the abrupt change in it gradient may cause it to deposit immediately the greater part of it load. In this way a *fan* is built up, radiating outward from the point at which the tributary emerges (Fig. 44).

The fan acquires its shape because of the nearly uniform distribution of rock waste over its surface by the stream that builds it. The stream repeatedly chokes up its channel, overflows, and forms new distributaries, just as in the braiding process. When one sector of the fan has been built up, the stream shifts to another and lower sector and builds that up. In this way the entire surface becomes covered by deposits laid down by the stream, and the fan as a whole is remarkably symmetrical.

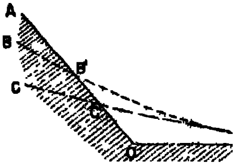


FIG. 45. Vertical section showing growth of a fan. Bedrock is shaded; stream deposits are stippled.

AOX = profile of surface before fan deposition.

$BB'X \doteq$ long profile of stream at an early stage of fan building.

$CC'X$ = long profile at a later stage, after the stream has become graded. In the grading process, the stream has cut away the original upper part of the fan, in addition to cutting away much bedrock upstream from the fan.

Upbuilding of the fan steepens the gradient of the tributary stream at the fan itself, while, upstream from the fan, erosion is in progress, reducing the gradient there. This dual action continues until the gradient has become just steep enough to enable the stream to transport its load down the fan. The radiating profiles of the fan thus become graded profiles (Fig. 45).

Fans are larger and more common in regions of high relief, at points where streams emerge from very steep slopes on to plains, than in regions of low relief and gentle slopes. In the mountainous parts of western North America some of the fans are several miles in radius and hundreds of feet in thickness. East of the Rocky Mountains, fans are less conspicuous and are confined chiefly to the bases of such steep slopes as valley sides laterally cut by streams.

Deltas. As the current at the mouth of a stream is checked by the standing water of the sea or a lake, the load drops to the bottom, gradually building an embankment that grows outward like a railroad or highway fill made by dumping. As the fill is built up close to the water surface, deposits usually encroach upon it from upstream, are distributed outward, and gradually cover it, making a crudely triangular area of new land with one apex pointing upstream. From this shape resembling the Greek letter Δ, the deposit derives its name of *delta* (Fig. 46). The term is used whether the embankment is converted into land or not.

The stratification of a delta is shown best in small deltas built in relatively deep water. Here deposition is most rapid on the outer slope, forming thick *foreset beds* sloping outward (Fig. 46). Fine sediment is carried farther out and forms thin *bottomset beds*. Along the top, where erosion and deposition alternate as the stream current fluctuates, thin *topset beds* are laid down. Some deltas are very thick; that of the River Po is at least 500 feet thick beneath the site of Venice.

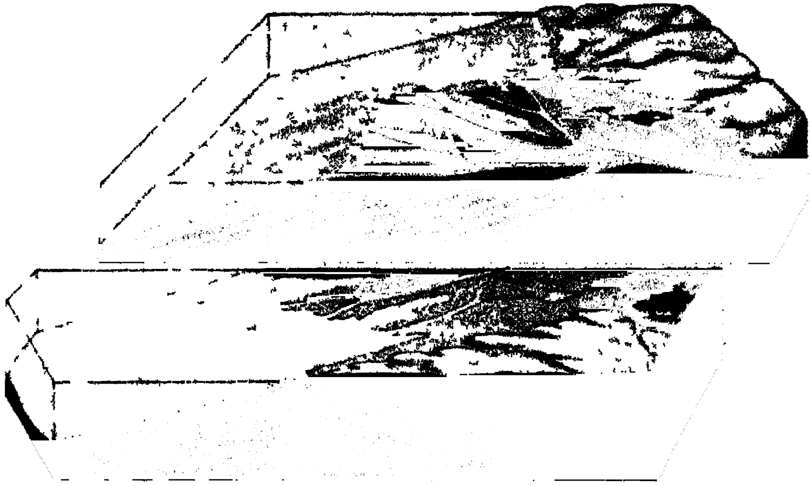


FIG. 46. Ideal small delta in three-dimensional view. On the surface appear "delta fingers" formed by deposition from distributary streams. The block has been cut in two and the pieces pulled apart to reveal a vertical section of the delta. Thick, inclined foreset beds are underlain by bottomsets and overlain by topsets. The topsets are represented only by the thickness of a single line, because they are commonly very thin. The slope made by the uppermost foreset bed is shown in phantom view, through the water. (Vertical scale greatly exaggerated.)

In the Mississippi delta and other similar deposits built by large streams into shallow water, the current is so strong and the water so shallow that this ideal arrangement of beds is not realized, and the stratification of the delta resembles the stratification of the alluvial deposits farther upstream.

A stream that rises in flood and overtops the natural levees in its downstream course breaches the levees at weak points, flows down their outer slopes (which are much steeper than the gradient of the channel), and continues onward through new channels, leaving the old channel with diminished discharge. New channels repeatedly formed make a branching system of distributaries that gives the delta its characteristic shape. Between the distributaries lie shallow basins which fill with fine sediment during floods and are thus converted into low land. The surface of the delta of a large stream so closely resembles the floodplain farther up the valley that the delta has to be delimited by an arbitrary boundary, usually the courses of the two distributaries farthest upstream.

The shifting of the main channel through the development of new distributaries is strikingly illustrated by the Hwang Ho (Fig. 47),

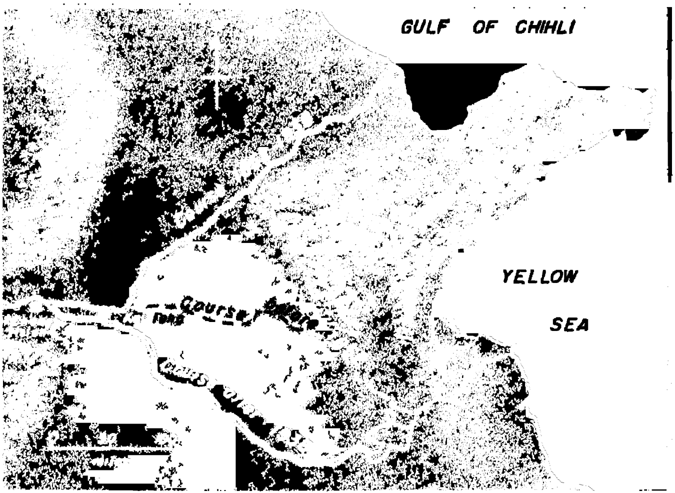


Fig. 47. Sketch map showing three distributary routes taken by the lower Hwang Ho in eastern China

which for approximately 700 years prior to 1852 had discharged eastward into the Yellow Sea. In an overbank stage in 1852 it spilled over at a point more than 300 miles above its mouth, formed a new channel northeastward down the floodplain, and finally emptied into the Gulf of Chihli, almost 300 miles north of its old mouth. The Hwang Ho occupied this new course, with minor distributaries, until 1938, when it again changed, this time through human agency. In that year Japanese forces at Kai Feng (Fig. 47) were moving on Ch'eng Hsien. The defending Chinese artificially breached the south levee of the Hwang Ho, diverting the river southeastward into an old distributary channel long abandoned and thus putting the stream between themselves and the Japanese. The invaders, their transport bogged down in the muddy flood, failed to reach Ch'eng Hsien.[1]

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[1] It was reported that late in 1947, through a great engineering effort, the river was once more diverted into its pre-1938 course

The Mississippi made a similar though temporary change in April, 1890, breaking its banks more than 100 miles upstream from its mouth, at a weak point caused by an irrigation ditch. From here the temporary distributary flowed east, entering the Gulf of Mexico 80 miles north of the main mouths, incidentally flooding a wide area, causing great damage, and halting railroad traffic for two months. As the flood subsided the river resumed its old course, after having taught local engineers that in the vicinity of the break-through the surface of the stream when normally bank-full is about *15 feet higher* than the flats beyond the natural levees. Even under normal conditions the Mississippi discharges a part of its water through the Atchafalaya River, a large distributary more than 150 miles in length.

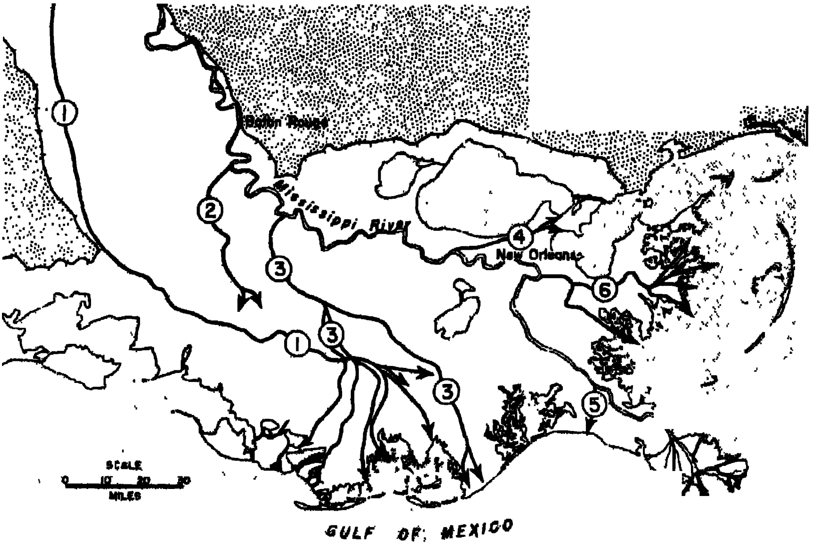


FIG. 48. The Mississippi delta and its surroundings, in Louisiana and Mississippi. White area represents delta and floodplain. Heavy-stippled areas are higher lands. The numbers refer to successive distributary routes used by the Mississippi River during the last several thousand years, as interpreted by R. J. Russell.

The Mississippi delta, one of the greatest in the world, with 12,000 square miles of land area, is in reality a complex of several coalescing deltas that have been built successively and continuously during the last several thousand years (Fig. 48). The building of each individual unit was begun at the time of some break-through and diversion that became permanent. Figure 48 shows the sequence of main distributaries inferred mostly from abandoned systems of natural levees.

CHARACTER OF STREAM DEPOSITS

Whatever their original shapes, stones and smaller particles transported by streams become rounded (Fig. 142, c, p. 206) through the abrasion they undergo as they are rolled and tumbled by the shifting currents. Furthermore, because of the delicate adjustment of transporting power to the sizes and weights of the rock fragments forming the load, stream deposits exhibit sorting according to size and weight (and therefore composition) of the rock fragments (p. 264). Layers of pebbles of various sizes are usually separated from layers of sand, and these in turn are sorted out from layers of silt and clay. No one layer extends far, however, because the variability of stream currents, repeatedly changing the conditions of deposition, results in thickening and thinning of the layers, and causes them to lie at angles inclined to each other (Fig. 121). The nearest approach to uniformity and parallelism of stratification is found in floodplain deposits and in the foreset and bottomset beds of deltas, all of which are deposited under conditions of minimum turbulence.

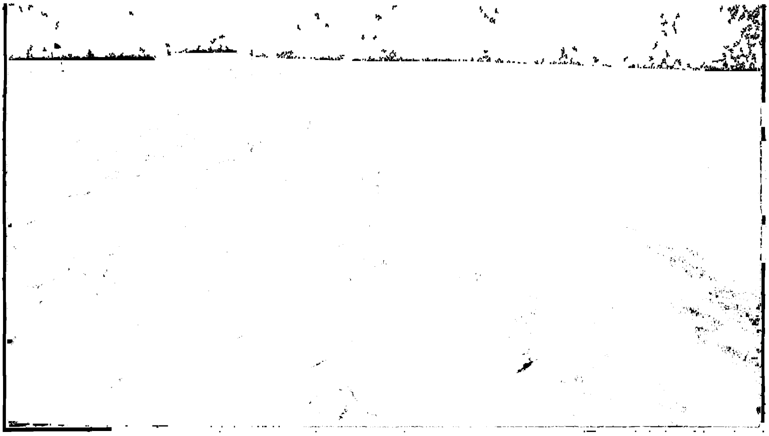
SOIL EROSION AND FLOODS AS ECONOMIC PROBLEMS

Soil Erosion. Of the agricultural land in the United States, nearly one-third has lost much of its topsoil, an additional third has lost all its topsoil, and one-eighth has been destroyed by erosion.* These are staggering figures, not only because of their size, but because they refer to the nation's basic natural resource—the soil. After two decades to two centuries of cultivation (depending on the location) man has destroyed much of what had been built up by natural processes through thousands of years. The big rivers had always carried sediment to the sea, but at a rate that was balanced by the slow creation of soil by weathering.

Then came the pioneer. He—and his successors—cut down forests, drained swamps, plowed the sod. The rain battered the ground unimpeded by the protecting leaves of trees and ran down bare slopes no longer checked by grass. The whole complex process of moving the rainfall to the sea was speeded up and made disastrously rapid, as increased velocity increased the capacity of the running water (Fig. 49).

But the damage does not end on the hillslopes. The great quantity of sediment washed into the streams buries agricultural bottomlands, silts up reservoirs, destroys fish life, and pollutes water supplies; and the too-rapid discharge of runoff, no longer adequately delayed by percolation through the ground, heightens floods on the big rivers to disastrous figures.

Soil erosion can be checked, and the normal regimen that existed before the damage began can be restored, by widespread adoption of conservation measures, among which are the following:



U. S. Soil Conservation Service.

FIG. 49 In this field near Bethany, Missouri, corn was planted in rows that ran downslope instead of along the contour. After a heavy rain gullies developed, 4 to 5 inches deep, through which the field as a whole lost an average of 2 inches of topsoil. Such damage can be prevented by proper methods of cultivation.

1. Retirement of marginal lands which should never have been deforested or plowed and which must be returned to their former condition without delay.

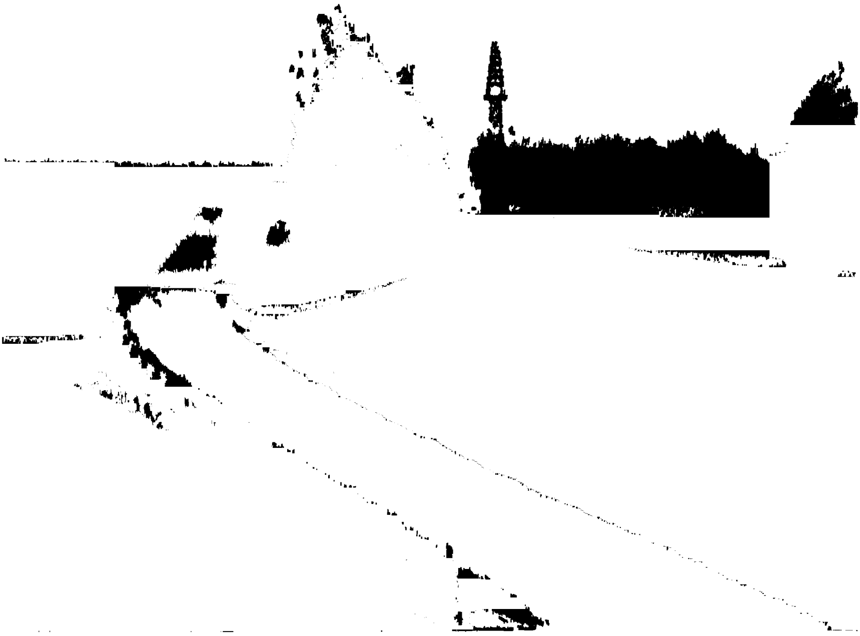
2. Soil-protection measures, aimed at slowing down the flow of water on slopes. These include

(*a*) Terracing the slopes.

(*b*) Plowing and cultivating "row crops" such as corn and cotton only *along* the contour.

(*c*) Planting strips of thick-growing crops (hay or grain) alternately with row crops along the contour, and staggering the seeding dates so that the entire slope is never bare.

(*d*) Rotation of thick-growing crops with row crops so as to permit small gullies to heal.



U. S. WEATHER BUREAU.

Fig. 50. Dynamiting an artificial levee on the flooded Mississippi River 18 miles below New Orleans, Louisiana, in order to give an additional outlet to the river to reduce the height of a 20-foot flood. River is shown at right; the road and lowland on the left were immediately inundated.

3. Curative measures, such as the healing of gullies by planting quick-growing vines and the building of check dams, and the proper draining of roads on slopes.

Destructive Floods. Prior to the settlement of North America, the regimen of the Mississippi and other large rivers included annual floods which were usually overbank, inundating broad areas beyond the immediate stream channels. Early in the eighteenth century attempts were made to confine the flooded Mississippi to its channel in order to protect new settlements and to reclaim for agricultural purposes the fertile floodplain. These attempts took the form of artificial levees (Fig. 50) which were built on the tops of the natural levees. If a given discharge is confined to a narrowed channel and thus prevented from normal flood-time spreading, the cross-section of the channel must be increased. We have seen that much of this added depth is attained locally by increased erosion of the bed, but part of it is attained by rise of the water surface. Hence, whenever a new levee is put in, those already in existence must be raised. The first levee, built at New Orleans, was 4 feet high. Today the average height is 18 feet. In 1902 there were 1300 miles of these levees along the Mississippi; in 1927 there were more than 1800. The river at flood stage is gradually rising higher above the adjacent basins, so that when breaks occur they are correspondingly more destructive. Witness the disastrous flood of 1927 in which the swollen river broke through the levees in 225 places and inundated 18,000 square miles of inhabited land.

Through more than 100 years each big flood on the Mississippi has been more disastrous than the last. It seems probable that this is largely the result of deforestation, artificial drainage, and improper agricultural practices throughout the huge region from the Appalachians to the Rockies that contributes water to the Mississippi. Even if such evils were not enormously harmful in themselves, the protection of the lower Mississippi basin would be sufficient reason for curbing them. Levees can not be built to withstand floods of ever-increasing magnitude, but probably they are capable of confining the normal floods which we could expect once again if drastic and far-reaching conservation measures were applied to the tributary region. In the meantime the extensive flood-control plan now under way recognizes the need of additional means of combating floods, among which are the building of dams on tributary streams, and the establishment of permanent overflow channels along old delta distributaries, to act as safety valves for regulating the height of flood stages.

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12. The more recent numbers in the series of Water-Supply Papers, published by the U. S. Geological Survey, contain a large amount of information on precipitation, runoff, stream discharge, and floods. See especially Water-Supply Paper 994, Cloudburst Floods in Utah, 1850-1938, published in 1946.

CHAPTER 6

SCULPTURE OF THE LAND BY STREAMS AND MASS-WASTING

Relation of Streams to Mass-Wasting. Having viewed the processes by which streams work, we may now examine the way in which streams sculpture a land area. At the outset we must realize that whereas streams are responsible for cutting and deepening their valleys, and for transporting the waste from the land as though they were vast systems of converging conduits, yet the shaping of the greater part of the land surface—including most of the side slopes of the valleys themselves—is the work not of streams but of mass-wasting (Fig. 51). To be sure, the cutting of valleys provides differences of height that permit gravity to operate on the mantle, and so mass-wasting is limited by the streams and is graded with respect to the stream profiles. But the slopes themselves are wasted by broad sheet-like movement of the mantle which, after working down to the base of a slope, is fed into the nearest stream. At the stream edge, the sheet-like movement is abruptly converted into linear movement as the streams sluice away the rock waste toward the sea or into inland basins.

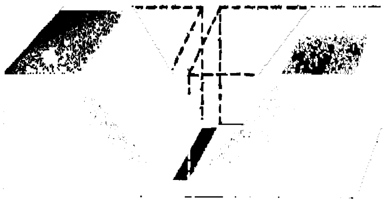


FIG. 51. Segment of an ideal valley. The narrow vertical cut shows the volume of rock directly excavated by the stream. The much greater remainder of the total excavation was accomplished by mass-wasting.

SCULPTURE IN MOIST REGIONS

DEVELOPMENT OF STREAM VALLEYS

Birth of Valleys; Consequent Streams. Some streams follow depressions made for them by other processes, but the majority occupy valleys that they themselves have excavated. Stream valleys begin their existence on new land surfaces just emerged from beneath lakes

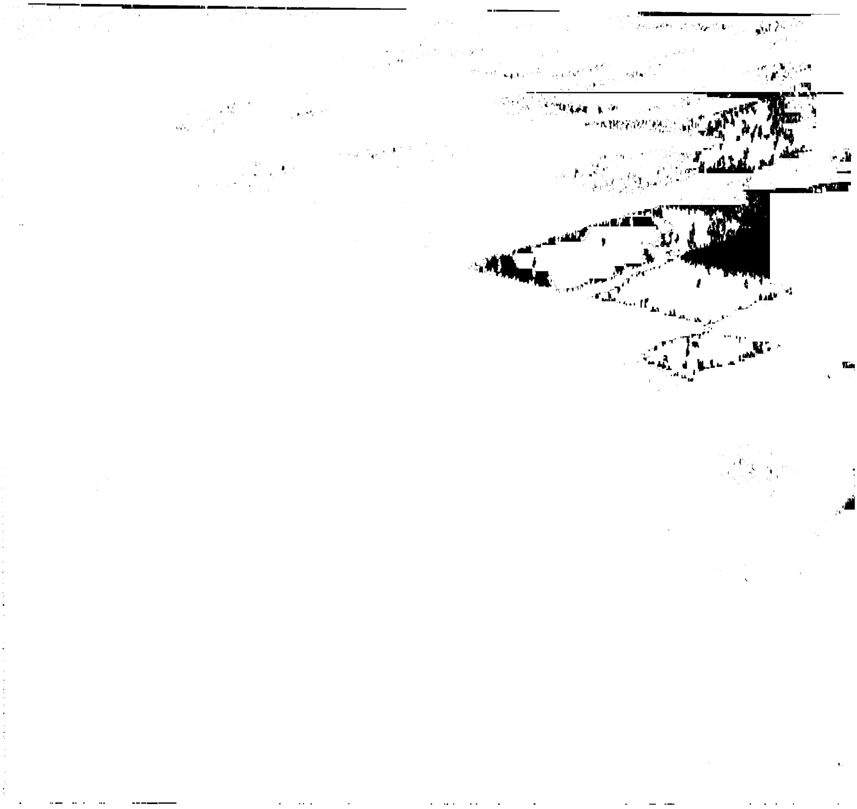

BRUBAKER AERIAL SURVEYS.

Fig. 52. Consequent system of gullies growing headward up a gentle slope underlain by silt. Horse Heaven Hills, north of Pendleton, Oregon. View from the air.

or the sea, or underlain by deposits recently made by glaciers, streams, volcanoes, and the wind. On new surfaces of all these kinds the first rains wash away loose particles and carry them down the nearest slopes in broad sheets and rills. Joining at the bases of converging slopes, two or more rills combine, and the increased velocity resulting from increased discharge causes increased erosion in more than direct proportion. With increasing erosion, the rills become larger (and of course fewer), and the larger they become, the more water they carry. Each rill first excavates a *gully* (Fig. 52). Erosion at its head, accelerated because the gradient is steepened there, causes the gully to lengthen headward, and the runoff washing down its sides widens it. At the same time the rill flowing through the gully deepens it. Thus undergoing continued enlargement, streams that follow (are "consequent upon") the pre-existing slopes of the land in the manner described are said to be *consequent* streams.

Intermittent and Perennial Streams. At a very early stage the valley is likely to carry water only during and after rains, so that its stream is intermittent. As erosion deepens the valley, however, more and more of the rainfall that enters the ground directly has opportunity to emerge again along the valley sides and to contribute ground-water runoff long after the rains have ceased. Eventually the valley is excavated to the depth at which all openings in the rocks are permanently filled with water. From this time on, water seeps steadily and uninterruptedly into the valley, forming a perennial stream.

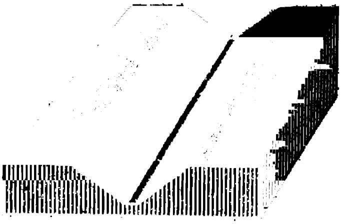


FIG. 53. Ideal main valley with sloping sides, showing why tributaries grow headward from it at acute angles. *oa* = slope of land surface; *ob* = slope of valley side; *oc* = resultant course of runoff.

Tributaries. Many of the tributaries of a consequent stream system are formed simultaneously with the main stream, by runoff down the existing slopes. Additional tributaries begin by flowing down slight irregularities in the sides of the main valley, cutting gullies whose heads grow upslope away from it (Fig. 53). Once begun in this way, the tributary gullies are extended headward at the same general angle to the main, because the growth of each is guided by adjacent tributaries developing under the same control. The pattern made by the whole branching system resembles the trunk and branches of a tree.

Divides. As a stream system continues to grow headward up the slope of the land, its headwaters approach those of another stream system draining in the opposite direction. The line separating the two watersheds is a *divide;* it is gradually lowered as the watersheds are reduced by erosion. At the same time it is shifted laterally, because no two stream systems perform equal erosion and the resulting greater rate of erosion on one slope shifts the divide slowly against the weaker system. The shift becomes increasingly slow, partly because gradients normally decrease as streams continue to erode the land, and partly because the rates of erosion on the two drainage areas gradually approach equality.

In both stream systems each tributary has a rate of erosion that differs from that of each immediate neighbor. Therefore the intervening divide is shifted unequally along various parts of its course, and so

SPENCE AIR PHOTOS.

Fig. 54. Land mass in early-mature stage of the fluvial cycle. Note the intricate maze of valleys and the zigzag pattern of some of the divides. Poso Creek district, Kern County, California.

it gradually assumes an irregularly zigzag pattern (Fig. 54). Where two opposed valleys head directly against each other, a sag develops in the divide.

Stages in the Evolution of a Valley. A valley once formed in a land area standing moderately above baselevel goes through a fairly definite series of changes, each of which merges with the next, to form an unbroken chain. In crude analogy with living organisms, valleys are said to be young, mature, and old, the characterization emphasizing their *stage of development* rather than their absolute age in years.

The ideal *young* valley has a V-shaped cross profile (Fig. 55) and its floor is scarcely wider than the channel of the stream it contains, because downcutting by the stream is dominant over valley widening by mass-wasting.

Irregularities in the channel and banks of the stream deflect the current, turning it against one bank. By impinging against this bank, it is turned toward the opposite bank, where the process is repeated. In this manner the channel takes on a winding pattern. The centrifugal force of the swinging current results in increased erosion of the banks at the places where the current is thrown against them. The banks are worn back at these places, which alternate, right and left, down the valley. As the banks yield, the channel shifts sidewise toward the outer and downstream sides of its curves (Fig. 56). At the inner and upstream sides of the curves the water is slack. Here alluvium is deposited and is abandoned as the channel shifts sidewise away from these areas.

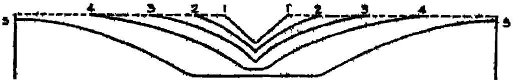


FIG. 55. Cross-profiles of a stream valley, developed successively during the progressive erosion of the valley. *1*, *2*, *3* are V-profiles. *4*, *5* are broadly flaring profiles.

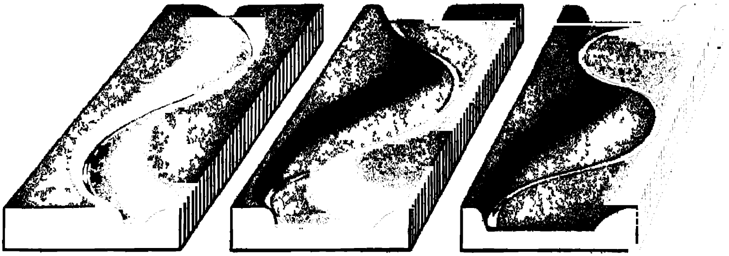


Fig. 56 Fig. 57 Fig. 58

FIGS. 56–58. Evolution of the valley floor and development of stream curves into meanders.

FIG. 56. Stream is actively cutting at the outer sides and slipping away from the inner sides of its curves, making crescent-shaped strips of valley floor.

FIG. 57. Later stage. Stream has cut a continuous strath and is undercutting the spurs that face upstream.

FIG. 58. Later stage. Valley has been widened nearly to the width of the curves (now meanders), which sweep down the valley.

The combination of downcutting and lateral shifting of the channel imparts to the valley sides alternating steep and gentle slopes (Figs 56, 57). The latter gradually merge into a slowly widening valley floor. At first the floor consists only of crescent-shaped strips on alternate sides of the channel (Fig. 56). But as the channel shifts and as downcutting diminishes, the strips coalesce, forming a continuous, more or less flat valley floor cut from the bedrock and thinly covered with stream deposits (Figs. 57, 58). This floor is known as a *strath*, a word long used in Scotland in reference to a wide flat valley floor. When the strath has become fully developed, the valley is said to be *mature*. By this time the stream, undercutting the valley sides, has widened the valley floor to a diameter greater than the diameters of the curves of the stream channel. When the curves have reached this condition they are termed *meanders* (Figs. 58, 59, 60). (Some meanders, however, are formed in other ways, as indicated on p. 85.) The steepest bank and deepest water lie on the outer side of a meander, where the current is strongest. As the curvature of a meander becomes more pronounced, a channel is cut through its neck, and an island is thus formed. The main current takes the shorter route afforded by the cutoff, and the entrances to the abandoned channel are gradually filled with sediment, leaving a shallow crescent-shaped (*oxbow*) lake (Fig. 59). Such lakes are undercut and destroyed by other meanders migrating gradually down the valley.

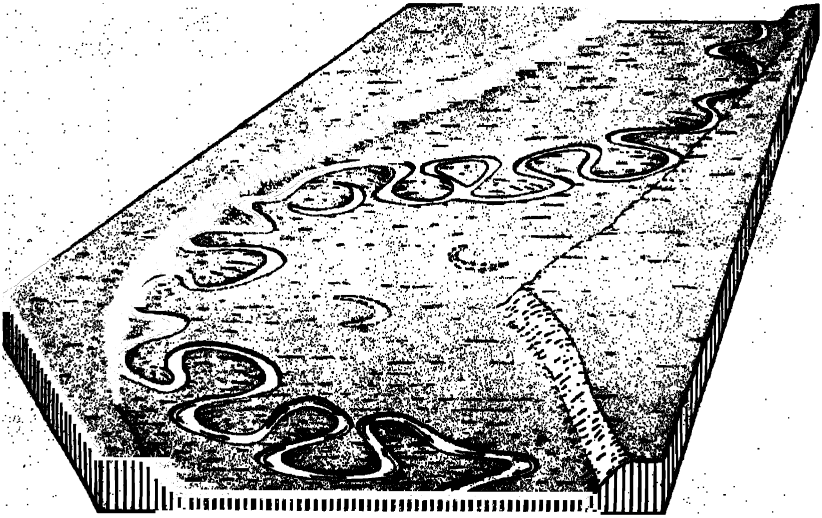
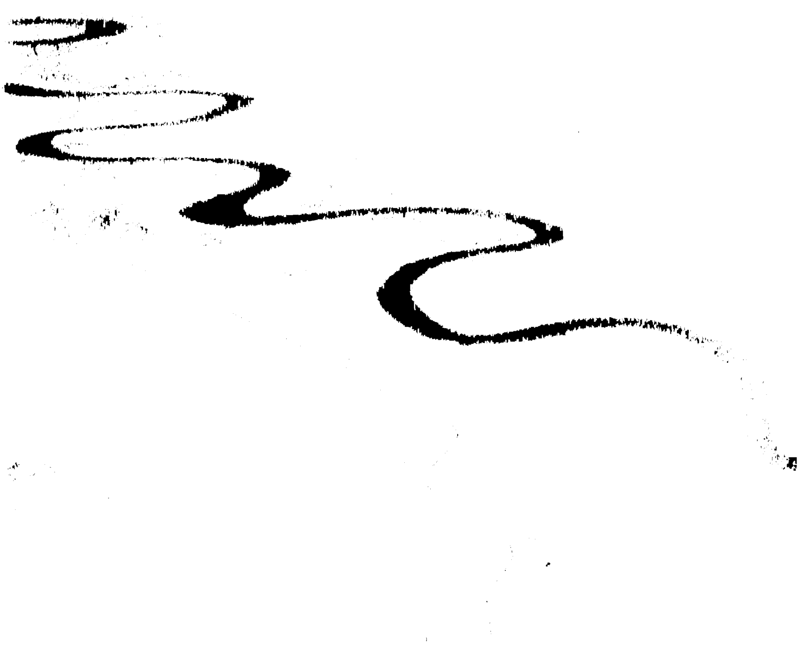


FIG. 59. Valley with wide strath, showing formation of cutoffs and crescent-shaped oxbow lakes. Arrows in near part of stream indicate course of current and position of deepest part of channel. Undercutting on outer banks and deposition on inner banks of the channel give the strath the faint interlocking slopes visible in the diagram.

Lateral cutting does not cease at this stage. The meandering stream impinges here and there against the valley sides, wearing them back still farther. The stream also shifts its course by overflow during



ROYAL CANADIAN AIR FORCE.

Fig. 60. Mature valley with wide strath and meandering stream. Annapolis Valley, Tupperville, Nova Scotia.

floods (p. 86). In these ways the strath may be broadened to many times the width of a single meander (and if the stream is subject to overbank floods, part or all of the strath may become a floodplain). Stream cutting takes place more and more slowly, and mass-wasting reduces the valley sides to very gentle slopes. The valley has now become *old.*

The terms young, mature, and old are convenient in describing valleys, for each term implies a group of distinct characters.

THE FLUVIAL CYCLE IN A MOIST CLIMATE

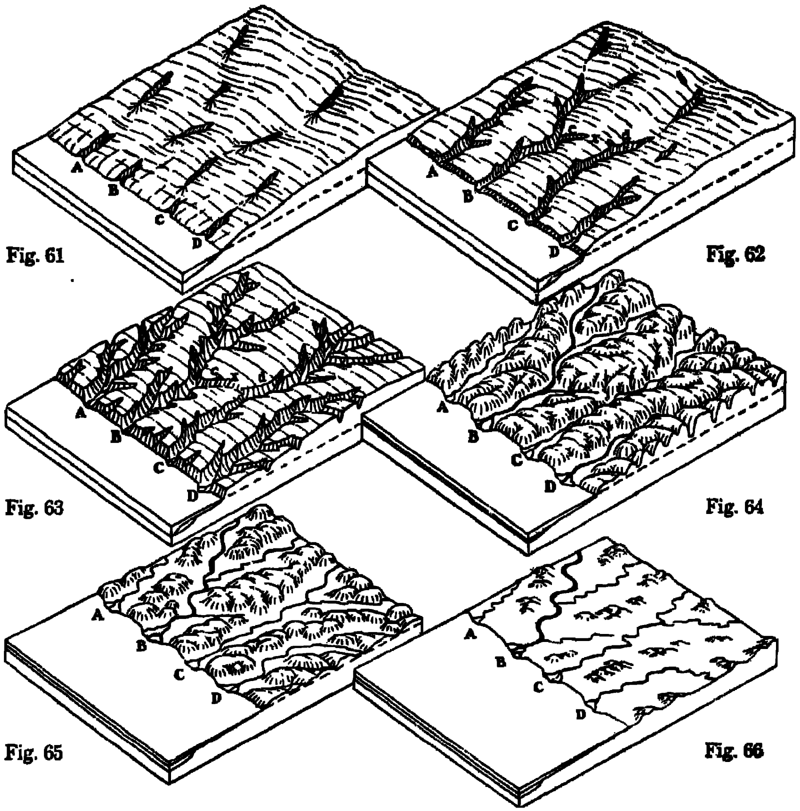
The change in size and form of a valley as it evolves through its young, mature, and old stages is marked. From visualizing the evolution of a single valley it is only a step to visualizing the evolution of a broad land mass drained by innumerable valleys. Since valleys enlarge at the expense of the areas that separate them, it follows that, given time enough, streams and mass-wasting must eat away even a great land mass and reduce it to baselevel. Such complete destruction is the certain result of these processes if they are not interrupted. The series of changes involved in the upheaval and the complete reduction of a region to baselevel, by any geologic process, constitutes a *geomorphic cycle* (a completed process involving the form of the land). The geomorphic cycle that is controlled by streams and mass-wasting is termed the *fluvial cycle.* Of course no eye has seen a single land mass undergo all the changes described, because the process is very slow. But because we observe many different land masses today, each in a different stage of evolution, and each grading imperceptibly into the next, we can safely infer that, given time enough and freedom from interruption, the entire sequence of changes will occur.

The fluvial cycle varies with climate, differing in the moist temperate climates of eastern North America and western Europe from the cycle in tropical regions with extreme rainfall, and again from the cycle in warm arid regions. We may select the first and the last as examples, because these are well represented throughout wide areas of North America.

Initial Stage. Figure 61 represents a land area composed of homogeneous rocks newly emerged from beneath the sea and lifted up to moderate height in a region of moderate rainfall. Gullies develop on its surface at once. Adjacent gullies join downslope and form connected chains, which gradually become valleys. Tributaries develop, lengthening themselves headward from the parent gullies, and the land area passes from the *initial stage* into *youth.*

Youth. The region is now drained by a stream system developed upon the pre-existing initial slopes, which happened to be so arranged that greater concentration of drainage took place along the lines of the infant streams *B* and *C* (Fig. 62) than along the streams *A* and *D*. Streams *B* and *C* therefore sent out more tributaries and grew headward more swiftly than their less favored weaker neighbors *A* and *D*.

The struggle for supremacy among these streams is illustrated by the situation at *x* (Fig. 62). Two streams, *c* and *d*, tributary respectively to *B* and *C*, have lengthened their valleys headward toward each other, narrowing the broad area that formerly separated them. Stream *c* has a shorter route to the sea than has stream *d;* hence the gradient of *c* is steeper, its power to erode is greater, and it cuts its valley both downward and headward more rapidly than does *d*. This inequality in rate of erosion results not only in lowering the divide *x* but also in shifting it away from *c* toward *d*.



FIGS. 61–66. Ideal fluvial cycle under a moist temperate climate and on homogeneous rocks.

FIG. 61. Initial stage, showing gullies developing at points where the runoff is concentrated.

FIG. 62. Early youth, showing growth of the stronger at the expense of the weaker streams.

FIG. 63. Later youth, showing reduction of the initial surface to irregular ridges.

FIG. 64. Early maturity, showing dissection of interstream areas into slopes and development of straths.

FIG. 65. Later maturity, showing decrease in relief, reduction of slopes, and widening of valleys.

FIG. 66. Old age, showing development of fluvial peneplane with monadnocks.

If the inequality were great enough, the divide would be shifted into the floor of stream *C* immediately below the former mouth of *d*, and all the drainage of *C* above this point would be suddenly *captured*, and diverted via *c* into *B*. Other types of stream capture are described on pages 115 and 498.

The streams in youth have gradients as steep as the height of the land above sealevel permits; they occupy V-valleys with steep sides and crooked courses. Tributaries develop rapidly, their valleys growing headward from the main streams like branches growing from the trunks of trees. As the numerous tributaries, powerfully aided by mass-wasting, dissect the surface, the broad areas between parallel streams contract into narrow irregular ridges (Fig. 63).

Maturity. When the tributaries have lengthened their valleys so far headward that the areas between them have been dissected into short hills and spurs, and when the growing valleys have consumed much of the land mass, the landscape imperceptibly takes on a new aspect and the region is said to be *mature* (Fig. 64). The intricate network of drainage is complete, and most of the area has been carved into slopes. The main streams have cut downward far enough to decrease their own gradients appreciably, and to develop straths on which they meander, while the upland (initial) surface, prominent during youth, has wasted into slopes. Because downcutting has been checked, the mantle moves down the slopes less rapidly. Sliderock accumulation and landsliding gradually give place to creep, and consequently the profiles of the valley sides are converted into sweeping curves (Fig. 70, *left*). The mass-wasted slopes are graded with respect to the streams at their bases, just as a mature tributary stream is graded to its main. In this way graded profiles extend not only along the streams but up the hillslopes as well.

Old Age. Because of the ever-decreasing gradients of the main streams, the heights along the steeper tributaries are now wasted more rapidly than the larger valleys can be deepened. The result is a gradual decrease in relief (Fig. 65). Lateral cutting by the sluggish main streams widens the valleys and helps to cut away the adjacent higher land. As the floors of the main valleys slowly approach baselevel, and downcutting by the main streams gradually diminishes, erosion is confined more and more to valley widening by the main streams and mass-wasting of the slopes, although downcutting may still be active along the lesser tributaries whose gradients are still considerable. The hills of maturity merge downward into low altitudes in old age, catching less precipitation and shedding the reduced runoff feebly. Tributaries diminish in number. Erosion takes place with increasing slowness, so that it might require more time to remove the last few

McLAUGHLIN AERIAL SURVEYS.

Fig. 67. Monadnocks near Moosehead Lake, Maine

et of land still above baselevel than was needed for the destruction all the land that went before.

Peneplanes; Monadnocks. The surface of low relief, very gently dulatory, that is developed in late old age is a terrestrial *peneplane* 'almost a plane," Fig. 66). Only a few residual remnants of the rmer high land remain. Here and there isolated hills of resistant ock stand somewhat above the general surface, like islands above the a (Fig. 67). Such island-like masses are *monadnocks,* so called after ount Monadnock in New Hampshire which stands in this manner ove the surface of the surrounding country. The creation of a ter- strial peneplane is the work of stream erosion and mass-wasting mbined, each process aiding the other.

If we followed the cycle to its theoretical conclusion, we should have add a stage in which the peneplane was converted by almost in- edibly slow degrees into a plane, devoid of elevations or depressions any kind. But, because we have never found remnants of such anes, we are inclined to doubt that they have ever been made, and we devote little attention to them.

The terms youth, maturity, and old age do not refer to periods of years, or to any absolute age. They denote merely stages defined by the amount of work done in proportion to the total amount of work that must be done to complete the cycle. A region of very weak rocks might reach old age while an area of resistant rocks was still in youth as far as the amount of erosion accomplished is concerned. It follows that an extensive valley system may be in different stages in different localities, depending on supply of water and on the nature of the underlying rocks. Other factors equal, main valleys are always more advanced in the cycle than are their tributaries.

Since streams have been carrying rock waste from land to sea for hundreds of millions of years, why has not all the land been reduced to sealevel? The reason is that the Earth's crust is very unstable; while one part is being worn down, another part is being lifted up and subjected to renewed erosion. For this reason there are no undamaged peneplanes in the lands of today. Remnants of former peneplanes exist, but none is now near baselevel. Some peneplanes have been lifted up, and dissected by a new generation of streams, so that they are now recognizable only as patches on flat-topped divides. Others have been buried beneath sediments of later date, and are now recognizable only where the blanketing deposits have been, still later, partly cut away: where the peneplane has been, so to speak, exhumed.

More often than not, a cycle once begun is interrupted, usually by renewed uplift of the land, before it has reached the peneplane stage. Results of such interruptions are discussed in Chapter 20.

Time Involved in the Cycle. The time required for the great sequence of changes involved in the cycle depends on the height of the land above sea, its distance from the sea, the resistance of the rocks to erosion, and the climate; but even under the most favorable conditions it is enormously long. From measurements of the amounts of sediment carried by the Mississippi River under various conditions, it has been computed that the vast land area tributary to this river is being lowered at the average rate of 1 foot in 7000 to 9000 years. This rate is very rapid in terms of geologic history; it seems rapid even in terms of human history when we realize that since the time when man first appeared on the Earth (about a million years ago) an average thickness of 100 to 150 feet of rock has been removed from the Mississippi basin. Yet at this rate more than 15 million years would be required to reduce the region to baselevel, even if no account is taken of the gradual slowing up of erosion as a result of decreased gradients. The actual time required would be vastly longer.

SCULPTURE IN ARID AND SEMIARID REGIONS

DIFFERENCES BETWEEN DRY AND MOIST REGIONS

Arid and Semiarid Regions Defined. Dry territory is usually classified roughly into *semiarid* (in which the annual precipitation is 10 to 20 inches, as in the Great Plains region east of the Rocky Mountains) and *arid* (which receive less than 10 inches of precipitation annually and are subject to great evaporation). Some authorities define arid regions as areas in which drainage does not reach the sea. On the latter basis, one-quarter of the land area of the globe is arid, if exception be allowed for long through-flowing streams such as the Nile and the Colorado. These streams maintain themselves through arid regions in spite of great evaporation and lack of many tributaries, because their headwaters in distant mountains give them a large and steady supply.

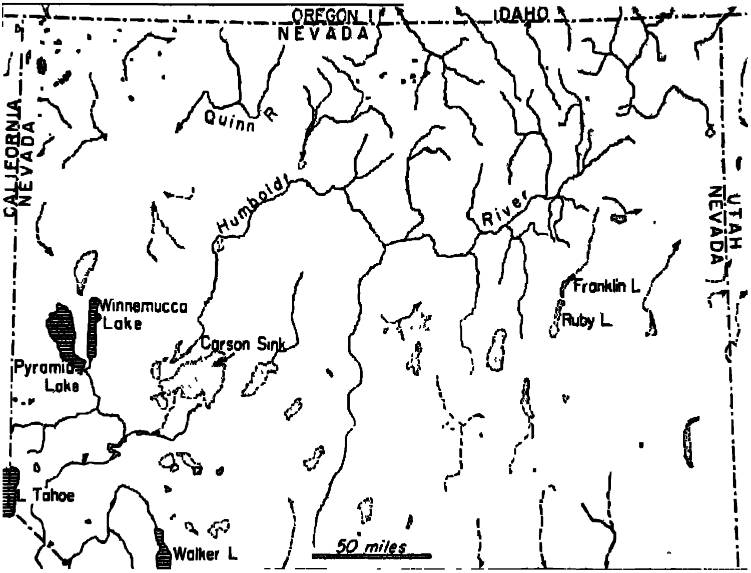


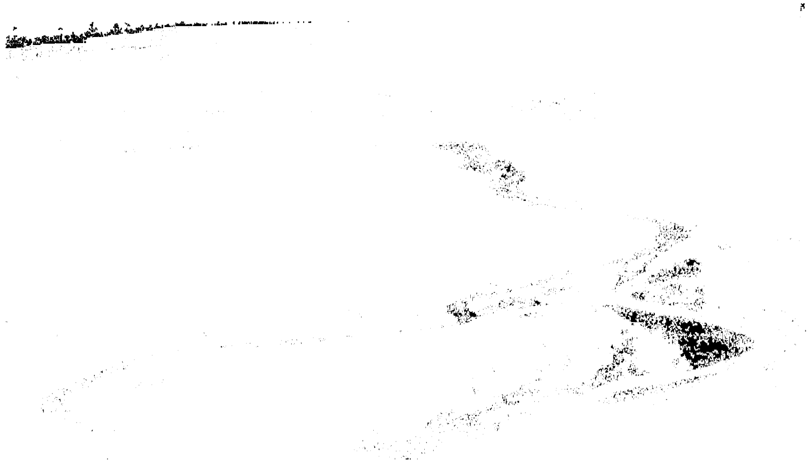
FIG. 68. Outline map of a large part of northern Nevada showing interior drainage. The long stream in the center is the Humboldt River; like the smaller streams it ends in an interior basin. Compare with the drainage of a moist region of somewhat similar size shown in Fig. 321, p. 489.

Chief among areas of interior drainage are the Sahara, the Libyan Desert, and the Kalahari in Africa, parts of the Basin-and-Range region in Nevada and adjacent States, the desert of western Australia, certain basins high in the Andes, and wide areas in central Asia, such as the Gobi Desert and the Takla Makan basin. All these regions are alike in that their streams lose themselves in the interior (Fig. 68).

Arid and semiarid climates favor some processes and inhibit others; as a result, sculpture of dry lands differs conspicuously from that of moist lands. The differences are explained below.

Weathering and Mass-Wasting. We have seen that in moist regions the mantle is nearly universal, comparatively fine textured, a product mainly of chemical weathering, and in motion downslope chiefly by soil creep. It is covered with an almost continuous growth of sod, brush, or forest. The resulting hill profiles, especially in the mature stage of the cycle, are a series of curves (Fig. 36).

In dry regions, on the other hand, the mantle is discontinuous, coarser in texture, and a product of mechanical weathering to a much greater degree. In the slow motion of the mantle down the slopes, talus creep replaces soil creep to a large extent. In consequence, slopes



SPENCE AIR PHOTOS.

Fig. 69. Butte and mesa along the Rio Grande at San Ildefonso, northwest of Santa Fe, New Mexico. The capping consists of lava flows resistant to weathering. These protect the weaker rocks beneath and maintain the cliffs near the summits. Probably the lava flows were formerly continuous through this district, the butte having been later isolated from the mesa through erosion by the Rio Grande and its tributaries

are generally steeper than in moist regions, and taluses are more commonly in evidence.

Mechanical weathering, working along joints in the bedrocks, forms angular blocks which, as they fall or creep downslope, leave steep rugged cliffs.

Hills made of flat-lying layers of rock remain flat topped and steep sided as they grow smaller in area through recession of the cliffs that constitute their sides. Flat-topped erosion remnants of this kind are *mesas*; similar features of smaller extent are the flat-topped *buttes* of western landscapes (Fig 69). The curving profiles that characterize the moist regions in maturity of the fluvial cycle are less common here.

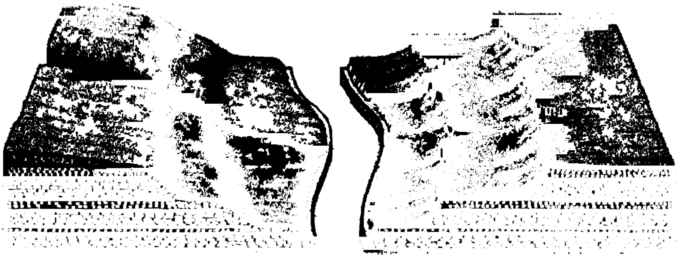


FIG 70. Contrast between landscape profiles in moist and dry regions. Left—resistant rocks largely masked by mantle of chemically weathered residuum creeping down the slopes Right—resistant rocks exposed as broad platforms and steep cliffs, with local taluses built by mechanical weathering and mass-wasting.

Mechanical weathering also accentuates instead of minimizing the effect of rock resistance on landscape profiles (Fig 70). No mat of creeping mantle obscures the contacts between weak and resistant rock. The Grand Canyon affords an excellent example It has been excavated by the Colorado River in nearly horizontal rocks arranged in alternating weak and resistant layers. The more resistant layers form broad platforms, and steep cliffs whose taluses (derived from mechanical weathering) only partly cover the weaker beds (Fig 70)

Stream Sculpture. In dry regions trees are generally very rare, and although in semiarid country a blanket of sod is present, the soil is loose, dry, and permeable. Gullying readily occurs where the soil is laid bare Because flow down all the slopes is little obstructed, the proportion of surface runoff to rainfall is relatively large, streams are subject to sudden floods, their waters are turbid, and in seasons of little rainfall their flow is greatly reduced and some go dry. The ability of the streams to erode, when in flood, is very great. In mountain areas the flood waters pick up great loads of rock debris. At the foot of the mountains stream velocities are abruptly checked, and discharges diminish not only by percolation downward into permeable stream deposits but also by evaporation enhanced by the dry climate. In consequence, fans are commonly larger and more conspicuous in dry than in moist regions. It should be noted that aridity aids fan development by promoting evaporation but is not essential to it. Mudflows help to build up the fans; they play a more important part in the progress of erosion and deposition than they do in moist regions.

Geological Survey of Canada]

FIG. 71. Badlands along Red Deer River south of Happy Jack Ferry, Alberta.

In many dry regions creep of the mantle down the sides of the smaller valleys is less conspicuous than the effects of lateral cutting and deposition by streams in flood. The result is a steep-sided, flat-floored valley or "box canyon," a characteristic feature of the Great Plains and other similar regions.

Although it is usual in dry territory for streams to be spaced far apart in response to the small precipitation, nevertheless weak fine-grained materials such as clay and silt exposed to erosion are rapidly carved into networks of gullies separated by narrow spurs. Such intricate gullying, which progresses visibly with every rain, is common along parts of the Missouri, Little Missouri, White, Cheyenne, and other rivers that drain the Great Plains region, most of which is underlain by weak bedrocks. Areas so gullied are termed *badlands* (Fig. 71).

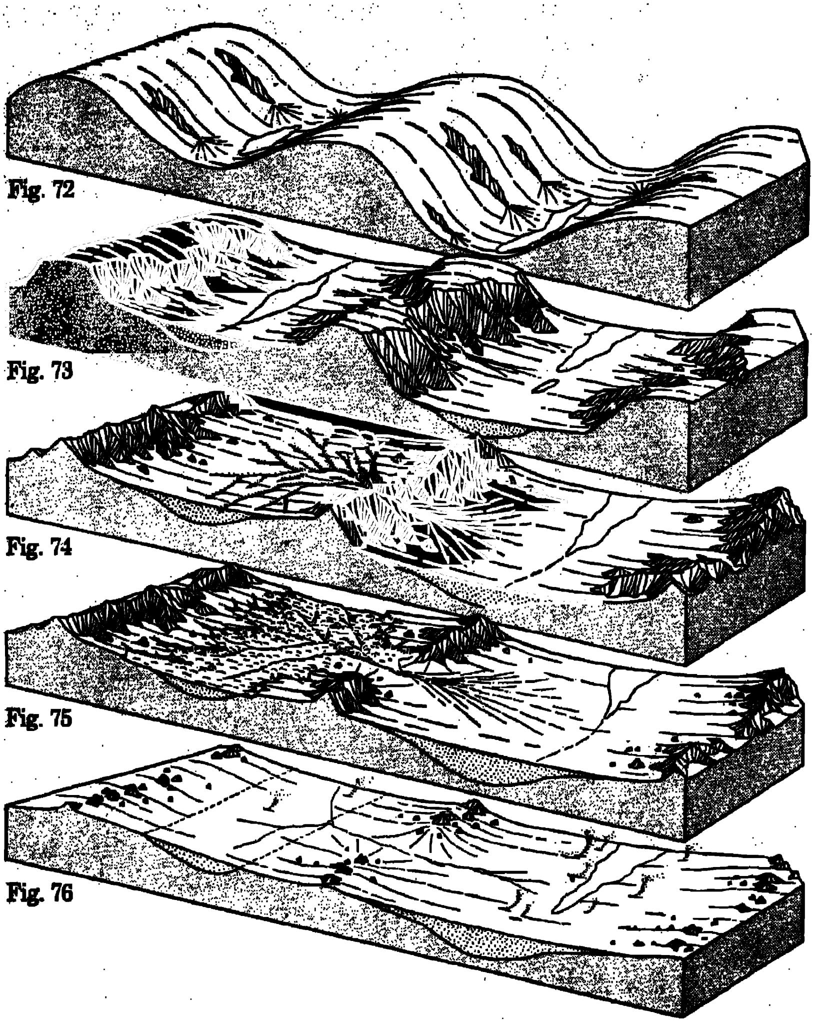
Influence of Wind. The small moisture content of the mantle, scanty vegetation, and rapid mechanical weathering that characterize dry regions are factors favorable to wind work (Chap. 10). In some deserts such as the Sahara, where sandstorms are frequent and violent, the wind plays an important part in shaping the landscape. However, wind work is not of prime importance in the arid regions of North America, because its effects are greatly exceeded by those of other processes.

THE FLUVIAL CYCLE IN A MODERATELY ARID CLIMATE

No single example can illustrate all the many kinds of arid regions. The idealized cycle outlined below represents the conditions embodied in the more arid parts of the southwestern United States, in which interior drainage is an important factor, but which is less dry than the Sahara.

Initial Stage. Under a sufficiently arid climate an initial land surface consisting of high ridges separated by broad trough-like closed basins results in interior drainage. As the ridges are uplifted, streams develop as consequents upon the newly formed slopes and carry drainage from the new highlands down into the adjacent troughs. They flow only after infrequent storms and are represented at other times only by dry valleys. Most of them evaporate or sink beneath the surface before they reach the low central parts of the troughs, depositing their loads in the form of fans built up along the bases of the highlands. The central parts of the basin floors are *playas* (Figs. 101, 102, p. 149); and the water that reaches them forms shallow lakes (p. 148) only to evaporate soon afterward. Figure 72 shows the initial surface, initial consequent valleys, and playas in the troughs.

Youth. In the steep highland valleys the streams are eroding their beds; on the gently sloping basin floors the same streams are depositing their loads. Where they emerge from highlands into basins they are at grade. Because they are here neither degrading nor aggrading, their energy is applied effectively to lateral cutting. Combined lateral erosion by several adjacent streams cuts a gently sloping surface in the bedrock at the base of the highland (Fig. 73). This surface is called a *pediment*, because of a resemblance, when viewed from a distance, to the triangular area that forms the gable of a roof. The making of pediments is a result also of weathering and mass-wasting on



Figs. 72–76. Ideal fluvial cycle under a moderately arid climate. (Compare Figs. 61–66.)

Fig. 72. Initial stage, showing ranges, basins, and early development of fans and playas.

Fig. 73. Youth, showing appearance of one type of pediments and decrease of relief as the basin floors rise by filling.

Fig. 74. Maturity, showing widening of pediments and capture of the higher basin by the lower.

Fig. 75. Later maturity, showing dissection of the higher basin, transfer of the waste to the lower, and the exposure of pediments.

Fig. 76. Old age, showing maximum development of pediments and disintegration of the drainage. Climax of deflation is indicated by small whirlwinds carrying dust.

the highland slopes and of other processes, the relative importance of each of which has not been established.

As the highlands are consumed by erosion their steep faces retreat, widening headward the zone in which the streams are at grade, and with it the pediment. Because the streams do not reach the sea, sea-level can exercise no control over them as it does over the streams in a moist climate. Here the baselevels that limit downcutting by the streams are formed by the basins themselves. The floor of each basin becomes the local baselevel for the streams tributary to it, and as the basins are slowly filled with waste from the mountains, *each baselevel rises*. This reduces the stream gradients and obliges the streams to reduce by ever-renewed lateral erosion the slopes of the pediments. At the same time the rising surface of the rock waste in the basins creeps higher up the pediments. The streams, still flowing intermittently but performing much work during their short periods of discharge, are at grade throughout the greater parts of their courses. Meanwhile strong winds pick up fine rock waste from the playas and sweep it, as dust, upward and outward over the inclosing highlands so that it escapes from the region entirely.

Maturity. As the mountain divide between two basins is cut down and the basin floors are built up, the higher basin in time is able to drain downward *across the old divide* into the lower basin. The drainage of the higher basin has been captured (p. 105) by the lower. When the drainage in the basins has thus become integrated into a unit the mature stage is said to have been reached, and drainage passes from the upper basin to the lower in times of rain (Fig. 74). The capture results in the dissection of the higher basin by a consequent system of gullies working headward from the new channel, and the waste from their excavation is deposited in the lower basin, hastening its filling. With the capture, the surface of the lower basin becomes the master baselevel. It controls both its own streams and those of its neighbor.

The dissection of the former playa in the upper basin makes badlands, and this basin becomes so thoroughly dissected that every part of it drains down into the master basin (Fig. 75), while the initial highland surface has entirely disappeared.

Old Age. With the lowering of the mountains the rains, infrequent at the outset, become even more rare because condensation decreases with decreasing elevation. The whole process is correspondingly retarded, but the wind, its erosive activity less interrupted by the occasional wetting of the remaining playa, blows away more rapidly the fine loose material of the basins. The pediment in the higher basin, stripped of most of its earlier mantle of waste by transfer to the lower, is laid bare. The wind becomes an increasingly important agent of erosion, picking up the finest material, blowing it away, and thereby slowly lowering the whole surface, which now resembles a plain more nearly than a pair of basins. Only the rock masses that are most resistant to mechanical weathering remain as monadnocks projecting island-like above the surface of the plain. Because the wind can work at will over the entire area, the surface continues to be slowly worn down (Fig. 76). In a very late stage the floor is thinly covered with coarse waste and dotted with monadnocks.

If the master basin were not closed, but instead drained out to the sea, the cycle would differ only in detail. Pediments would form in the master basin, but its floor would not rise by filling.

If the final stage does not exist in any arid region at present, it may be partly because the crust has been unstable or because insufficient time has elapsed since the glacial ages (when most dry regions were far less dry than now) to allow it to develop. It has been argued, however, that in the arid cycle the surface could be worn down by deflation (p. 200) to a depth well below sealevel, providing the sea were kept out by surrounding highlands. An example is the Qattara depression in the Libyan Desert in northeastern Africa. Nearly 150 miles long and in places lying more than 400 feet below sealevel, this basin was the southern anchor of the "El Alamein line" during World War II.

The wind, however, can not pick up rock particles from wet ground, and so ultimately deflation would be stopped by the water table, below which the rock is kept permanently moist. The water table (p. 120) therefore may constitute the baselevel for the cycle in such a region.

The cycle in a region such as the Sahara or southern Arabia differs from the above chiefly in a different emphasis on the processes. In those regions the initial relief was generally less, and the making of fans and pediments is less conspicuous relative to wind work than is the case in the United States.

CONCLUSION

Given any kind of land mass acted upon by streams under specified climatic conditions, we are able to foresee the general sequence of events that will take place and thus to reconstruct the past and predict the future by a study of the present. The organization of our knowledge of the work of streams into a continuous chain such as is represented by the cycle concept is an invaluable aid in the study of the sculpture of the land by streams.

The examples of land sculpture given in this chapter are based on simple conditions such as homogeneous rocks and a quiescent crust. The sculpture of many land masses has been more complex, but discussion of them is deferred to Chapter 20. This chapter follows the discussion of sedimentary rocks and crustal movements, a prerequisite to an understanding of the wide variety of stream-sculptured forms.

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Chapter 7

SUBSURFACE WATER

One of the reasons why man has been able to establish permanent settlements, not only in well-watered country but also in desert lands, is that there are few districts in which holes, intelligently located and sunk far enough into the ground, do not encounter at least a small supply of water. The depth of an adequate well in a moist country may be only a few feet; in a desert it may have to be hundreds of feet. The Earth's superficial rocks constitute a vast reservoir of *subsurface water*, which not only is of prime human importance, but also plays an important part in the endless process of geologic change. Water in the ground dissolves some of the substances of which the rocks are composed and thus prepares them for attack by other eroding agents. Conversely, it acts as a transporting and depositing medium for substances that cement loose sediments to form new rocks. The quantity and distribution of water beneath the surface control the flow of rivers, the levels of lakes, and the locations of swamps. And, finally, subsurface water furnishes the water supply of a large proportion of the world's population.

Source, Distribution, and Movement

Source and Distribution. Most of the water below ground is merely rain and melted snow that has seeped downward from the surface. Once it has penetrated into the ground, it follows various courses. Part of it finds its way back to the surface as springs or through the pores in the mantle or through plants; some of it is held within pore spaces in the rocks, and some of it actually becomes a part of the rocks, by uniting chemically with them.

In most regions with moist climate, wells generally yield water at slight depths. However, most mines and other holes that penetrate to depths of several thousand feet encounter little water at such depths; in fact many are dry. Furthermore, at several miles' depth the pressure of the overlying rocks is so great that open spaces capable of containing water probably can not exist (p. 445). It is likely therefore that the Earth's crust contains free water only throughout a restricted zone near the surface. We say "free water" because at greater depths the water present is an integral component of the rocks themselves, chemically united to them, and is therefore not available for withdrawal.

Within the zone in which subsurface water occurs, the chief factors that control its distribution are climate and the character of the rocks. Climate, through precipitation and evaporation, governs the amount of water contributed to the ground. Rock character governs the amount the rocks will absorb.

Porosity. The amount of water that rocks can contain depends on the open spaces in the rocks. The open space consists of very small openings, such as those between the grains of which the rock is composed, as well as of larger openings such as fissures and caverns. The *porosity* of a mass of rock (or mantle) is determined by the proportion of open space it contains. The porosity of some igneous rocks is less than 1 per cent (of the volume of the rock), whereas the porosity of some sands and gravels is 25 to 45 per cent. If the sands and gravels have become consolidated (p. 260) so that they form sandstones and conglomerates, their porosities may be reduced by more than half.

The ability of a rock body to hold water is increased by the presence of fissures and caverns, which add to the open space in a rock *body* but are independent of the porosity of the rock itself. Fissures are important in making a rock body more permeable, as discussed below.

Permeability. *Permeability* is the capacity of rock or mantle to permit water to pass through it, and in porous rocks it varies roughly as the square of the diameter of the particles of the material. Where, as is common, particles of several sizes are present, the smallest rather than the largest sizes determine the permeability. Thus gravels and sands ordinarily yield the greatest amounts of water, whereas silts and clays yield less. The porosities of some clays exceed 50 per cent; yet even when saturated these materials are impermeable, because the pores are so minute that the water is held by the molecular attraction of the clay particles. On the other hand, a compact granular rock such as granite, although practically without pore space, is permeable if it contains fractures, because water flows easily through the fractures.

SUBSURFACE WATER IN HOMOGENEOUS PERMEABLE ROCKS

Water Table Separating Vadose Water from Ground Water. In most regions the rocks are saturated with water below a depth that depends largely on the permeability of the rocks, the amount of rain water that sinks into them, and the topography of the land. In permeable rocks this surface below which the rocks are saturated is the *water table*. The subsurface water that lies below the water table is *ground water*; that which lies between the water table and the Earth's surface is termed *vadose water*. In the vadose zone the water content fluctuates rapidly, and generally the available openings are very incompletely filled.

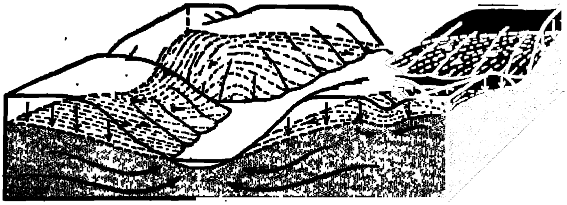


FIG. 77. Ideal diagram showing the relation of the water table to the land surface, the movement of the subsurface water (indicated by arrows), and the zones of vadose water and ground water.

For simplicity, the area shown is regarded as underlain by homogeneous permeable rock. The ground above the water table is drawn as if it were transparent, to show more clearly that the water table is a subdued replica of the land surface. Note that the ground water tends to move toward the stream valleys, where part of it emerges.

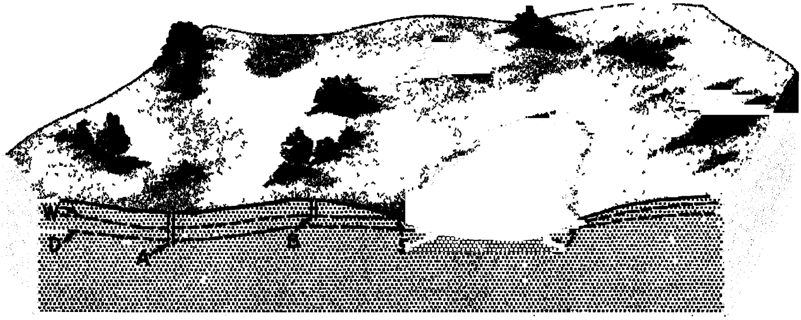


FIG. 78. Conditions controlling locations of springs and wells in homogeneous permeable rocks or mantle. The water table fluctuates between *W* (in the wet season) and *D* (in the dry season). Well *A*, reaching below the dry-season water table, has a continuous supply. Well *B*, reaching the water table only during the wet season, goes dry at times. *S*,*S*, zone of springs whose migration, with the fluctuating water table, causes the swamp to contract and expand seasonally.

The water table is not a "level"; on the contrary it is irregular. In moist regions and where the rocks are homogeneous, it is a subdued replica of the land surface beneath which it lies (Figs. 77, 78), standing relatively high beneath hills and declining toward valleys. The vadose zone is thickest beneath hills and becomes thinner toward valleys; in fact in many low places the water table intersects the surface, forming springs and merging with streams, swamps, and lakes (Fig. 79). The water table has gradients analogous to the gradients of streams. These gradients ordinarily slope down from points beneath hills to points beneath valleys. The gradient in any place is determined by the permeability of the rock and the quantity of water supplied, usually from the rainfall immediately above. Thus in times of rain the water table rises and becomes more irregular, whereas in times of drought it subsides and flattens out. In general, the lowest position of the water table in any region is closely controlled by the vertical position of the largest surface stream, which acts as a kind of baselevel for the water table. As streams gradually lower their valleys by erosion, the water table of the whole region keeps pace by slowly subsiding, until in the peneplane it lies scarcely above the level of the sea.

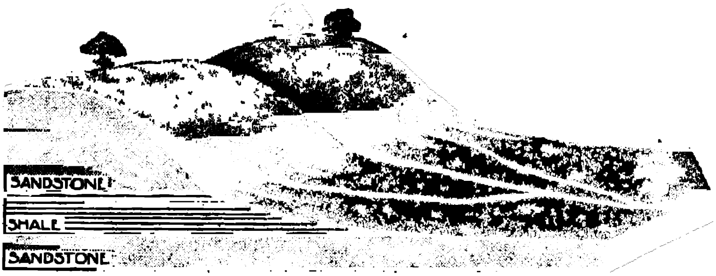


FIG. 79. Springs emerging along a contact between permeable sandstone and underlying impermeable shale. The springs are fed by water that has percolated down through the sandstone, and give rise to small surface streams.

In many dry regions the water table is highest beneath large streams flowing from well-watered mountains and is kept high in such regions by percolation from the streams themselves.

Movement of Subsurface Water in Homogeneous Permeable Rocks. Subsurface water moves chiefly by percolation. The movement of the vadose water is for the most part directly downward (Fig. 77). The ground water (the water below the water table) moves in the same direction as the surface runoff, but more slowly, because it encounters friction in its passage through the interstices of the rock. It moves most rapidly beneath stream channels in sandy and gravelly deposits that cover the bedrock floors of many valleys. In gravel the subsurface discharge (known as *underflow*) may be considerable, constituting a vast body of slowly percolating water beneath the bed of the surface stream. Under favorable conditions, underflow velocity may amount to at least several feet per day.

The continuous contribution to springs, streams, and lakes from this great underground reservoir is the cause of the continuous flow of streams in moist regions.

In general, the movement of ground water decreases as depth increases, because of loss of gradient and diminishing permeability. At the greatest depths to which ground water can penetrate, movement virtually ceases.

Simple Springs and Wells. Subsurface water emerges naturally at the surface either as unconcentrated *seepage,* or with a distinct current as a *spring.* There are many kinds of springs. Two common kinds of simple springs are illustrated here. Figure 78 shows the conditions under which springs and seepages occur along the trace of the intersection of the water table with the surface of the ground. Figure 79 shows a type of spring, localized at the contact of permeable rock on impermeable rock. Such a spring might consist of either vadose water or ground water.

A well (*A*, Fig. 78) excavated below the lowest position of the water table is assured of a continuous supply of water, replenishing itself automatically as water is drawn out (provided, of course, that amount of water and permeability are adequate). The rate at which water is withdrawn, however, has to be adjusted to the permeability of the rock, because wells in very permeable rock replenish themselves much more rapidly than do wells in rock of low permeability. A well (*B*, Fig. 78) that does not reach the water table at all times will intermittently become dry.

Wells are an important source of public water supply. In the United States alone, about 6500 public waterworks are supplied from wells, many of which are simple wells like those described.

SUBSURFACE WATER IN HETEROGENEOUS ROCKS

In areas underlain by two or more kinds of rock of differing permeability, movement of the subsurface water is necessarily more complex. Of the many possible conditions three principal ones stand out: perched water bodies, artesian systems, and fracture systems.

Perched Water Bodies. If impermeable layers are present, descending water is stopped at their upper surfaces. In areas where the water table lies well below the surface, an irregularly shaped mass of impermeable rock in the zone of vadose water may hold an isolated body of ground water perched or suspended above the normal ground-water zone. The isolated body has its own local water table—a *perched water table*—below which shallow wells can obtain water (*W*, Fig. 80). Conversely, impermeable strata may extend below the regional water table. A hole penetrating such a stratum would be dry even after it had passed below the general level of the water table in the surrounding region.

Confined Ground Water: Artesian Systems. In a series of inclined sedimentary rocks that include a permeable layer such as a sandstone,

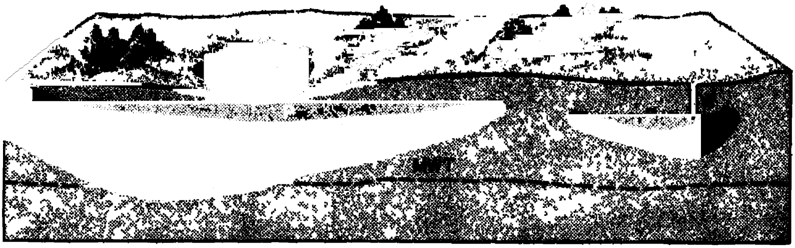


FIG 80. Perched water bodies (*P*, *P*) caused by irregular impermeable beds (shown by dark shading) lying above the main water table (*MWT*) and yielding water to a shallow well (*W*). A deeper hole (*H*) ends in impermeable rock and fails to obtain water.

overlain by an impermeable layer such as a shale, subsurface water moves freely in the sandstone but is held immobile in the shale. Free hydraulic connection between the ground water in the sandstone and the ground water lying above the sandstone is prevented by the impermeable shale except where the sandstone is cut by the surface of the ground. Here, at the outcrop of the sandstone, water enters this layer. Its percolation is *confined* to the sandstone by the overlying shale, which forms a "roof," and thus there is developed a hydraulic gradient distinct from the slope of the water table of unconfined ground water. As the shale itself is saturated, the whole system lies below the water table. The permeable layer or *aquifer* may be tapped by drilling a hole through the confining impermeable layer. Then the hydrostatic pressure of the column of water extending up to the intake of the aquifer forces the water up through the hole, to a height equal to the vertical distance between intake and outlet, minus a small amount owing to friction resulting from percolation through the aquifer. If the hydrostatic pressure is sufficient to lift the water in the outlet hole above the water table, the hole is an *artesian well* (Fig. 81), so called because the first well of this kind was made in the French province of Artois in the twelfth century. Some true artesian wells flow out at the surface, but many have to be pumped. Artesian wells can not be made simply by boring deeply unless the requisite conditions stated above are present. Deep wells bored into rock so as to intersect the water table (some of them penetrating far below it) are often called artesian wells, but this is an incorrect use of the term; there is no difference in principle between wells of this kind and ordinary shallow wells.

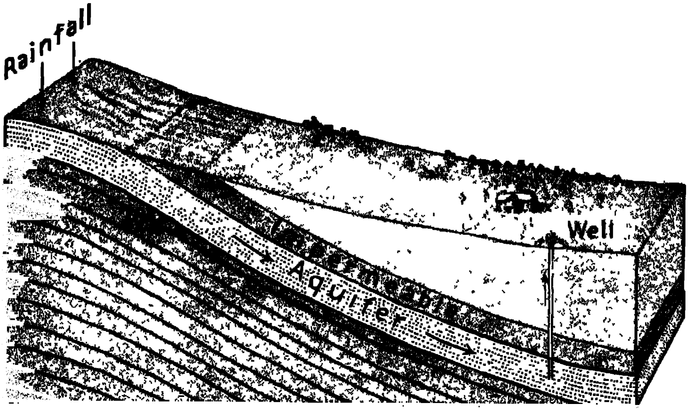


FIG. 81. A simple artesian system. A permeable sandstone layer (an aquifer) crops out in the hill to the left and absorbs rainfall which descends along the sandstone, sealed in beneath a capping layer of impermeable rock. A hole bored down into the aquifer, farther down its slope, becomes an artesian well. Near the well, the water table may lie in any position above the aquifer.

In a sense an artesian system is like an inverted perched water body. In the latter, an impermeable mass prevents the descent of downward-percolating water. In the artesian system the impermeable mass prevents the upward movement of water under hydrostatic pressure.

Dakota Artesian System. One of the most important artesian systems in the United States is the Dakota system. The aquifer is the Dakota sandstone, with a porosity of 10 to 20 per cent and a high permeability. It crops out and absorbs water along the flanks of the Rocky Mountains, Big Horn Mountains, and Black Hills, and underlies large parts of Kansas, Nebraska, the Dakotas, Wyoming, Montana, and Saskatchewan. The first well was drilled into this aquifer in 1882; since that time more than 15,000 have been drilled. The resulting withdrawal of water has seriously reduced the hydrostatic pressure, and the yields of all the wells have diminished, and are still declining, indicating that the natural supply of water is still being overdrawn. Some of the States concerned have resorted to legislative action to make conservation compulsory. This will hasten the arrival of the time when withdrawal will balance supply.

Another important system is the Illinois-Wisconsin system. Here there are two aquifers, the Potsdam sandstone and the St. Peter sandstone. They crop out in Wisconsin and descend southward, becoming gradually less permeable in this direction owing to increase in clay content. Velocity of percolation in the aquifers has been calculated as about one-third mile per year. The beds of partially indurated sand that underlie the Atlantic Coastal Plain from Long Island to Texas provide artesian systems of considerable local importance. Brooklyn, New York, and neighboring Long Island communities derive large supplies from such artesian sources. The public water supplies of Memphis, Tennessee, and Houston, Texas, are furnished by artesian flow from sandstone aquifers.

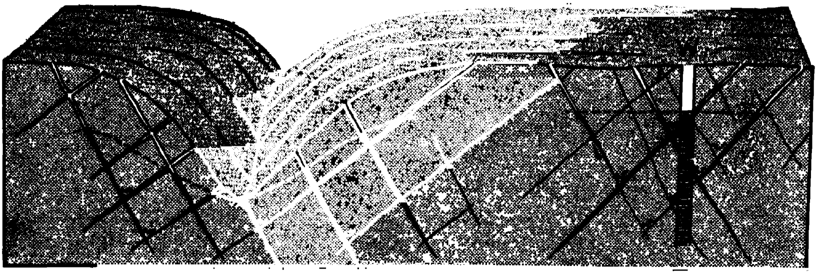
In some districts wells must be bored very deep before artesian water is encountered. In Berlin, St. Louis, and Pittsburgh, for example, the necessary depth is about 4000 feet. Along the Atlantic coast, on the other hand, most of the artesian wells are only 100 to 300 feet in depth. The volume of water developed by some is large; the great 12-inch well of St. Augustine, Florida, with a depth of 1400 feet, supplies 10,000,000 gallons a day.

Although the largest artesian systems are dependent on beds of sandstone, smaller but none the less important supplies of confined water are derived from lenses of sand and gravel in fans, glacial deposits, and other superficial accumulations.

Artesian Springs. Artesian systems operate also where the confined ground water has a natural outlet (usually along a fissure) instead of a well; the resulting outflows are *artesian springs.* Such springs have more uniform flow and are less affected by droughts than are the springs of unconfined water illustrated in Figs. 79 and 80. The water is generally cold, but the water from some deep aquifers, having passed through the higher temperatures that prevail at depth (p. 442), becomes heated. This is probably the explanation of such thermal springs as those at Hot Springs, Virginia. In some regions the water, while in the rocks, dissolves unusually large amounts of mineral matter and gives rise to "mineral springs," as at Saratoga, New York; Hot Springs, Arkansas; Carlsbad in Czechoslovakia; Bath in England; Wiesbaden in Germany; and Vichy in France. Strictly speaking, all springs are mineral springs, because all contain mineral matter in solution, but the term is applied in general to springs that differ markedly from ordinary water in the quantity or character of the mineral matter dissolved.

Thermal springs include both warm and hot springs. The latter, together with geysers (intermittently eruptive hot springs), are characteristic of the last dying phase of volcanism. Ordinarily they occur in regions of recently extinct volcanoes such as the Yellowstone Park district (Chap. 14).

Springs and Wells in Fractured Rocks. Nearly all igneous and metamorphic rocks have small porosity and permeability, and hence the circulation of water in them is restricted largely to cracks. Water enters the openings from above and moves down them in the vadose zone. In the ground-water zone it is confined, and under hydrostatic pressure it moves down or up the joints and fissures, according to local conditions (Fig. 82). In places it becomes imprisoned in cracks with no outlets; in others it seeps out at the surface.



Modified after U. S. Geological Survey and C. F. Tolman.

FIG. 82. Subsurface water in rock in which circulation is confined chiefly to fractures. Arrows show directions of circulation. Water table (dashed line) is controlled by depth of master valley. Well (*W*) is fed through fractures. Because fractures pinch out downward, the best water supply is obtained from moderate rather than great depths.

As holes put down in rocks of this kind derive appreciable flows of water only from fissures, the success of drilling depends largely on chance intersection with fissures. One well may furnish an abundant yield while another, a few score feet away, may yield only a trickle. In general it is unwise to drill deep if a good yield is not forthcoming from a moderate depth, because openings of all kinds decrease with increasing depth.

Although the majority of wells producing from fractures in crystalline rocks are of the normal type, some are artesian. Few of the latter, however, produce without pumping, and their yields are likely to be less than yields from rocks of other kinds. In consequence cities situated in areas of igneous and metamorphic rocks usually have to rely on surface water (in many cases artificially constructed reservoirs) for their public supplies. New York, the New England cities, Dublin, and Vienna are examples.

POLLUTION AND SANITATION

The most common source of contamination of water in springs and wells is sewage, and the infection most commonly communicated by polluted water is typhoid. Drainage from cesspools, broken sewers, privies, and pigpens percolates readily into the ground and contaminates the subsurface water. Although slow percolation through the ground purifies contaminated water, the purification process requires percolation through 100 or 200 feet at least, and in many cases very much more. Hence shallow wells must be located carefully with regard to possible sources of pollution, and even deep wells must be cased or otherwise sealed near their heads where there is danger of contaminated surface water's leaking into them (Fig. 83).

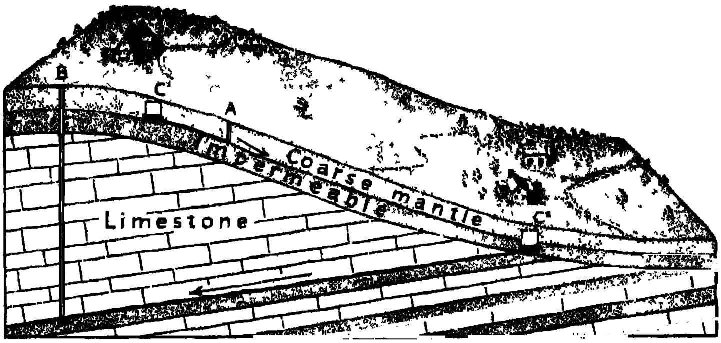


FIG. 83. Contamination of wells. The shallow dug well *A* was unwisely located downslope from a cesspool C^1 and therefore received contaminated drainage from it. The owner then drilled a deeper well at *B*. This well tapped layers of soluble limestone inclined toward it from the direction of the lower cesspool C^2. The water dissolved openings in the limestone and flowed, unpurified by slow percolation, to where it was drawn out through well *B*. The owner of the two wells will have either to relocate his cesspool or to dig a shallow well somewhere near *B*.

GEOLOGIC WORK OF SUBSURFACE WATER

SUBSURFACE WATER AS A SOLVENT

Subsurface water, like surface water, effects hydraulic action, abrasion, solution, transport, and deposition. However, its hydraulic action and abrasion are effective only in the restricted localities where there are underground streams (p. 136). As the chief factor in adding to the weight and mobility of mantle on slopes, it is an important mechanical element in the processes of mass-wasting (Chap. 4). But the greatest work of subsurface water lies in solution, transport of the dissolved substances, and deposition by precipitation. So widespread are these activities that the soluble mineral matter in the rocks of the continents is being slowly eaten away and delivered to the streams by the subsurface water. Containing oxygen, carbon dioxide, and acids acquired during its descent below ground, the water is an efficient solvent. It dissolves the cementing substances between the grains of resistant sandstone, causing the rock to break down into loose sand. It decomposes the feldspars and iron-bearing minerals in igneous rocks, destroying the interlocking-crystalline texture of the rock and making it crumble away. Thus it is the chief factor in the chemical weathering of rocks (Chap. 3).

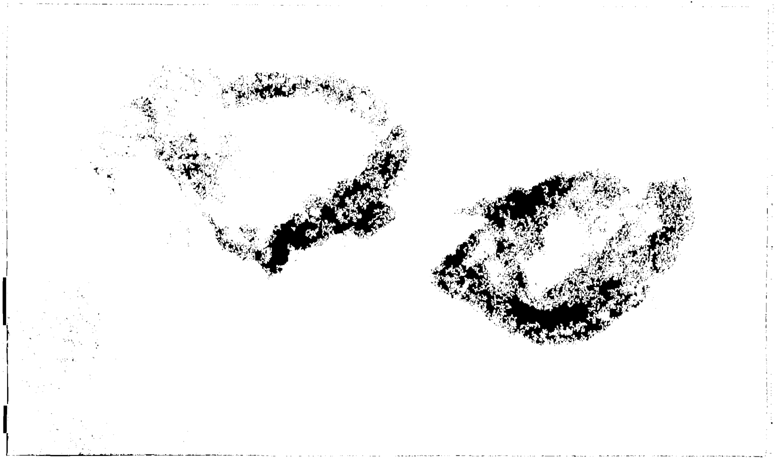
Some of the dissolved matter is reprecipitated below ground, as detailed in the following discussion. Another part is carried off by streams and poured into saline lakes and the sea. By evaporation of the lakes it is precipitated as salts of various kinds, but a large part of the dissolved matter carried to the sea has remained in solution, making the sea more salty year by year. In certain parts of the sea calcium carbonate is precipitated from solution to form beds of limestone (p. 258). The substances precipitated, whether in the sea, in lakes, or below ground, help to build up new rocks which in the course of time are dissolved again.

MINERAL DEPOSITS

The most important causes of precipitation of mineral matter from solution in subsurface water are (1) evaporation, at the surface, in caverns, and in minute pores, (2) loss of carbon dioxide, (3) reduction of temperature, (4) reduction of pressure, (5) chemical reaction between the mineralized water and the rock through which it is passing, and (6) action of minute plants (*algae*).

Deposits in Pores: Cementation. Percolating water precipitates mineral matter in the pores of rocks and mantle, thus reducing the porosities of rocks and gradually converting loose and unconsolidated sediments into sedimentary rocks (Chap. 12). Calcite, silica, iron compounds, and other substances dissolved from the rocks at and near the surface are carried away and precipitated as a cement at lower levels and in other regions.

Deposits in Larger Openings. Open spaces larger than pores are very commonly filled with mineral matter precipitated from solution in subsurface water, usually below the water table. A cavity filled with crystals (such as quartz or calcite), with the crystals pointing inward toward the center of the cavity and only partly filling it, is called a *geode* (Fig. 84). Concentric deposits of amorphous silica occur in a similar manner. One explanation of such features is that cavities formed by solution are filled at a later time, owing to a change in physical-chemical conditions, with mineral substances precipitated from subsurface water.



Ward's Natural Science Establishment.

Fig 84 Geode, resulting from the partial filling of a cavity with crystals of quartz Erosion has removed the inclosing rock (a limestone in southwestern Illinois), but the resistant quartz that lines the cavity has remained This geode (actual diameters 4½ x 8 inches) has been broken open to show its interior

Deposits in Caverns. Precipitation by subsurface water is most clearly evident in the deposits of calcite in the form of *dripstone,* with which many caverns are ornamented Dripstone is formed only in caverns that are above the water table and are therefore filled with air. Vadose water charged with calcium carbonate percolates downward from the surface of the ground to the roof of a cavern, where, clinging to the ceiling, it forms droplets. It evaporates a little, loses some carbon dioxide, and therefore deposits some calcium carbonate. Then, as more water is added from above, it falls to the floor below, where more of it evaporates, leaving another minute deposit. As the drops slowly succeed each other, long "icicles" of calcite (*stalactites*) grow downward from the roof, while broader accumulations (*stalagmites*) grow upward from the floor. If the process goes on long enough, each pair coalesces and forms a column. Dripstone likewise develops in irregular sheets on roof and floor, and assumes many fantastic shapes curious to the cavern visitor, but all these deposits are formed in this way (Fig. 85).

W. T. Lee, U. S. Geological Survey.

FIG. 85 One of the "rooms" in Carlsbad Cavern, New Mexico, showing dripstone deposits. Stalactites are forming at points where water emerges along cracks in the roof, broad-based stalagmites are growing upward, and some stalactite-stalagmite pairs have grown together to form massive columns.

Replacement. In some places the water dissolves matter already present and concurrently deposits an equal volume of another substance that it was carrying in solution. In this way a tree trunk buried below the water table may become slowly converted into petrified wood (Fig. 86) by the dissolving of its woody matter and the simultaneous deposition of silica. The fact that such *replacements* take place volume for volume is shown by the amazingly complete preservation of the fine texture of the original material. Many ore bodies owe their origin to replacement processes (p. 532).

Concretions. The capacity of subsurface water to dissolve and redeposit is well illustrated by the formation of certain types of *con-*

cretions—oddly shaped nodules that occur in sedimentary rocks. These forms, gradually built by precipitation around definite nuclei, are described on page 274.

Spring Deposits. Confined water rising from deep sources contains mineral matter in solution, and, emerging as springs (usually thermal), it precipitates some of the dissolved substances. Among the factors causing precipitation are relief of pressure, evaporation, and the action of minute organisms in the water. The substance most commonly deposited by springs is calcium carbonate, because it is the most soluble of the common rock-making substances. The physical character of the deposit depends chiefly on the rate of deposition. Deposits formed by slow evaporation are firm and compact. The dripstones formed in caverns are of this type. Calcium carbonate formed by rapid deposition from thermal springs, on the other hand, is likely to be porous and spongy. This less compact material is *travertine.* It is precipitated in mounds and terraces, some of which have picturesque forms. The terraces at Mammoth Hot Springs, in Yellowstone Park, are a well-known example.

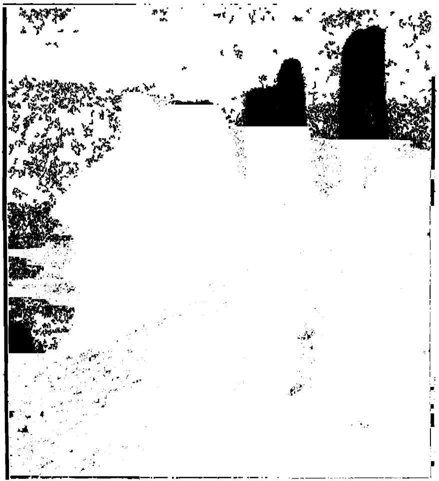

J. P. Iddings, U. S. Geological Survey.

FIG. 86. Tree trunks petrified by replacement while buried in volcanic ash and lava flows, in the positions in which they grew. The petrified trunks have since been exposed as a result of erosion of the inclosing volcanic material.

Other mineral substances are deposited in the same way. Even silica, ordinarily difficult to dissolve, is precipitated by some thermal springs, as in Yellowstone Park, forming *siliceous sinter*. Limonite, iron carbonate, sulphur, and gypsum are other substances deposited by springs.

SOLUTION IN CARBONATE ROCKS

Limestone (calcium carbonate), dolomite (calcium-magnesium carbonate), and marble (a soluble metamorphic rock of similar composition) constitute a group of carbonate rocks that underlie many millions of square miles of the Earth's surface. They are soluble in water charged with carbon dioxide (p. 33), and therefore in moist regions these rocks are vigorously attacked by subsurface water with striking results.

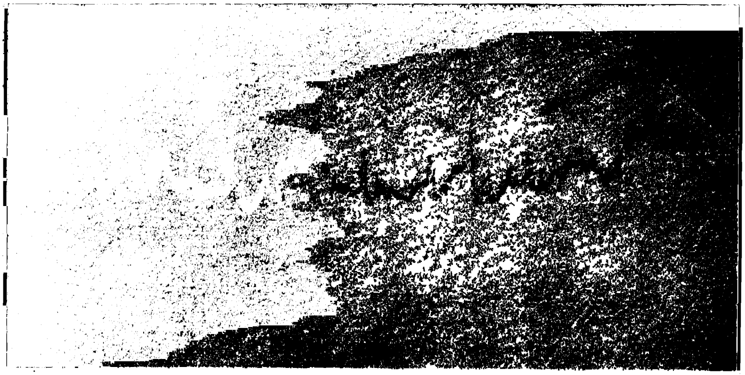


Fig. 87. Stylolite seam in Indiana limestone. The seam lies parallel with the stratification of the rock. The rock face on which the seam appears has been artificially cut. Length of seam, 12 inches.

Stylolites. The importance of solution in destroying carbonate rocks under favorable climatic conditions is shown by *stylolites*, which consist of a series of interlocking small columns of rock, intertoothed like the bristles of two hair brushes forced together. As the rows of columns lie in more or less horizontal belts, they appear on vertical faces of rock as zigzag lines (*stylolite seams*, Fig. 87). The way in which stylolites form is not fully understood. They may be the result of irregular solution, under pressure, at surfaces that separate two beds of soluble rock. Again, they may result from the differential compaction of the limy sediment during the slow process of its conversion into solid rock. General solution in the zone of weathering results in the more rapid lowering of the ground surface on carbonate rocks than on adjacent less soluble rocks. Some lowlands in moist regions have been made in this way. Carbonate rocks, however, are less subject to reduction by mechanical processes than are many other kinds of rocks. In consequence, in arid regions, where solution is at a minimum, carbonate rocks form cliffs and mountains that stand out conspicuously (p. 41).

Sinks. In compact, well-stratified, and strongly jointed carbonate rocks, the avenues of easiest descent for vadose water are vertical joints and planes of stratification. Those avenues most favorably situated with respect to supply from above and free circulation below are readily enlarged by solution as the descending water passes through them. Enlargement is most effective at the surface, where the water moves most rapidly and where it is freshly charged with carbon dioxide from the atmosphere and from decaying vegetation, and decreases rapidly downward. In consequence, the point of intersection of two joints near the surface becomes a funnel-shaped depression. As the depression widens, the overlying mat of insoluble mantle and vegetation collapses into it, and a *sink* is formed (Fig. 88).

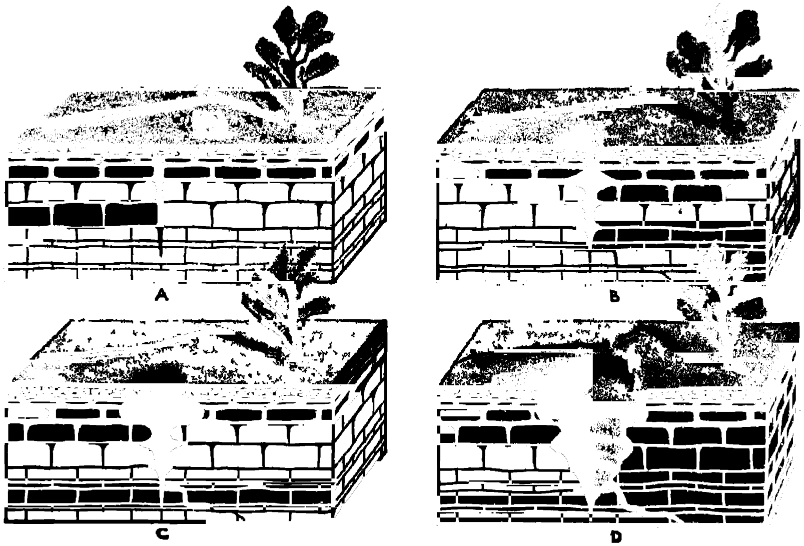


FIG. 88. Development of a funnel sink in compact, jointed limestone overlain by a thin residual mantle of insoluble weathered material covered with sod. The water table lies approximately at the base of the block. Length of front of block, 150 feet.

A, enlargement of a vertical joint by solution. *B*, irregular widening of the enlarged joint chiefly by solution along stratification planes. *C*, continued enlargement, most effective near the surface where the solvent capacity of the descending water is greatest. *D*, collapse of the insoluble mantle, forming a funnel sink.

Probably the majority of sinks are of this (funnel) type, excavated above the water table. Not uncommonly they form definite patterns on the surface, controlled by the structural planes along which they develop. They range in size from small openings only a few inches in diameter to great depressions hundreds of feet wide. Many are remarkably symmetrical, whereas others exhibit irregularities resulting from differences in the composition and structure of the rocks.

Another type of sink forms in massive, permeable carbonate rocks little affected by definite planes of weakness. Under these conditions solution is less sharply localized and occurs irregularly throughout considerable volumes of rock, probably both above and below the water table. Solution cavities of irregular shape are gradually excavated (Fig. 89) and enlarged. Eventually their unsupported roofs collapse, forming great depressions in the surface. These, like the funnel-shaped forms, are termed sinks, but differ from them in their asymmetry, in the debris of the collapsed roofs with which their floors are strewn, and in a sagging of the beds of rock around their rims, testifying further to the failure of their roofs. Some sinks of this type are very large. One near Mammoth Cave, Kentucky, is said to have an area of 5 square miles. Probably this size has been attained by the coalescence of a number of adjacent caverns or sinks.

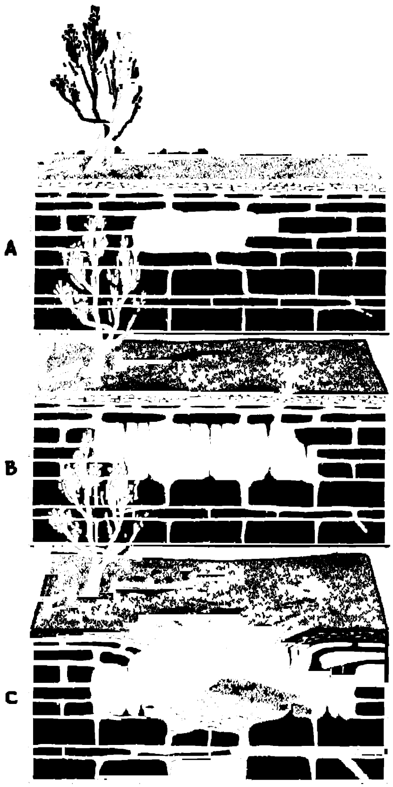
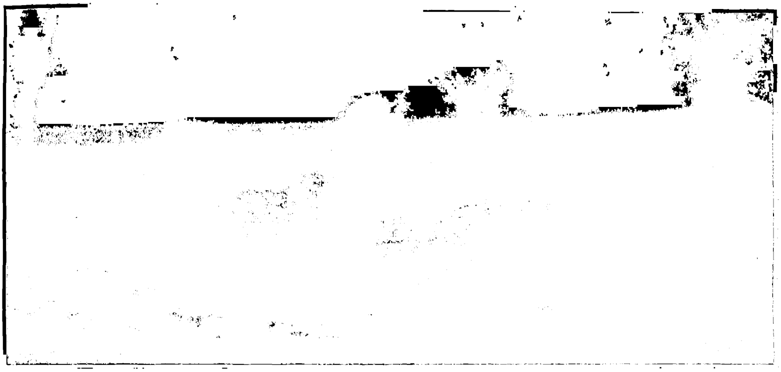


FIG. 89. Development of a sink of irregular shape through collapse of a roof. The water table lies approximately at the base of block.

A, formation of a small cavern in the more soluble strata of a limestone series. *B*, enlargement of the cavern and development of dripstone deposits in the form of stalactites and stalagmites. *C*, collapse of the cavern roof, forming an irregular sink. The debris of the roof litters the sink floor.

Sinks excavated below the water table usually contain lakes that persist as long as the water table is not lowered below the floors of the sinks. Most of the small lakes of northern Florida are of this kind.

On the other hand, the funnel sinks, having been excavated above the water table, drain downward through openings in their floors and therefore are usually dry. The outlets of some, however, are clogged by clay, humus, and other insoluble matter washed into them, allowing the development of small lakes whose levels are above the water table and are independent of it. In some sinks the water leaks away slowly; in others the insoluble stopper is suddenly broken through and the lake disappears with a rush.



I. D. Scott.

FIG. 90. Sink in limestone near Sunken Lake, Presque Isle County, Michigan.

Caverns. Caverns of many shapes and sizes occur in carbonate rocks in all parts of the world. Notable among them are the caverns of the limestone districts of Kentucky, Tennessee, and southern Indiana, the Shenandoah Valley in Virginia, northern Florida, Cuba, Yucatan, parts of the Philippines and Indo-China, the Karst region in Yugoslavia, and the Causses of southern France. Although most caverns are of small size, there are notable exceptions. The Carlsbad Caverns in southeastern New Mexico (Fig. 85) extend down to a depth of more than 1300 feet and include one chamber 4000 feet long, 625 feet wide, and 350 feet high. A single passage in Mammoth Cave, Kentucky, is said to be more than 8 miles in length, and individual chambers reach heights of 80 feet and widths of 250 feet. Certain vertical shafts in the Causses of southern France are more than 300 feet deep. Some caverns are linked up in connecting networks of galleries and shafts underlying areas of many square miles. Mammoth Cave has 30 miles of continuous passages.

The only true underground streams occur in cavern systems. Elsewhere the flow of subsurface water is chiefly by slow percolation through the rocks. Some cavern streams have considerable discharge, have widened their subterranean conduits somewhat by abrasion, and have deposited thin layers of clay on their floors. Such mechanical erosion and deposition, however, are insignificant in comparison with the solution and precipitation that have chiefly fashioned the caverns.

Some caverns are empty, whereas others are ornamented with dripstone. The latter record a change from conditions favoring solution to conditions favoring precipitation. Probably the reason for this change is twofold: (1) Some caverns were excavated below the water table, and, later, as the water table subsided during dissection of the land, they were drained and thereby made ready for replenishment with dripstone by vadose water. (2) Other caverns, however, appear to have been excavated above the water table and replenished with dripstone shortly afterward, or even while excavation was still in progress, by vadose water which was evaporating and losing carbon dioxide during the process of dripping.

Caverns and underground streams in the vadose zone are closely related to the regional system of surface streams. The underground streams, through the medium of the water table, are controlled vertically by the profiles of the master surface streams, fluctuate with the seasons, and gradually dissolve out caverns deeper beneath the surface, as the surface streams deepen their valley floors throughout the fluvial cycle.

Many "caves" are merely rock shelters in cliffs. They are generally made by differential weathering at the surface and are not true caverns. In past times, both rock shelters and caverns often served as refuges for primitive man and as the dens of animals now extinct Because of this the bones of men and animals, stone implements, and other objects have accumulated in them and have been sealed up beneath deposits of calcium carbonate and other substances slowly accumulating on their floors. Relics of this kind, especially in certain parts of Europe, have revealed much concerning the life and culture of prehistoric times.

Solution Valleys. The growth of a series of closely spaced sinks connected by a subterranean stream channel (Fig. 93) leaves a series of unreduced areas between the sinks. Enlargement of the sinks narrows the areas between them and reduces them to natural bridges, like the famous Natural Bridge near Lexington, Virginia, 150 feet high. As the bridges are undermined, an open valley is formed. Although they may have a general resemblance to valleys excavated by surface streams, these *solution valleys* are recognized by their highly variable width, by their originating (and in some places ending) in steep-walled sinks, by natural bridges that may still span them, and by tributary valleys left hanging above them through having lost their water to subsurface drainage.

Karst Topography. In some districts, sinks, caverns, and solution valleys are so numerous that they combine to form a peculiar topography with many surface depressions, unsystematic drainage patterns, and disappearing streams. This type of landscape has been termed *karst topography* (Fig. 93) because it is strikingly developed in the Karst region of Italy and Yugoslavia. It is developed also in such other regions as central Tennessee, Kentucky, and southern Indiana, northern Florida, Cuba, Yucatan, and the Causses of southern France.

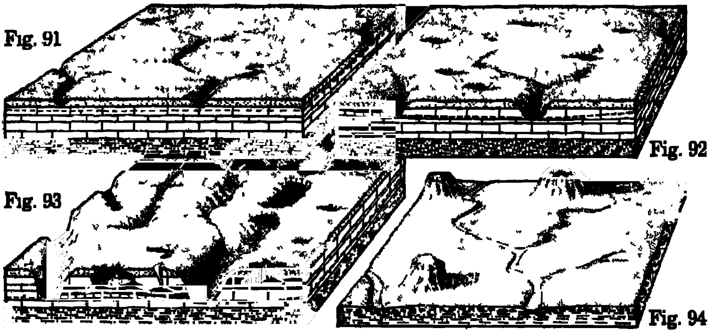
The Geomorphic Cycle in Carbonate Rocks. The evolution of the surface through solvent erosion involves a cycle parallel with the cycle of erosion by mass-wasting and surface streams. The following description of an ideal cycle is based on the assumption that the effective solvent action occurs above the water table.

Initial Stage. Assume a region underlain by thick horizontal beds of compact, strongly jointed limestone resting upon impermeable shale, both limestone and shale lying well above sealevel. The limestone has been protected by a thin impermeable capping such as well-cemented sandstone, which, however, is now stripped away at two or three points, laying bare the limestone surface. The region is drained by surface streams, and the water table stands at a considerable depth below the surface (Fig. 91).

Youth. At those points where jointed limestone is exposed at the surface, sinks of the funnel type begin to develop (Fig. 92). The water from them finds its way along a network of joints and stratification planes to one of the major valleys, where it reappears at the surface.

The water draining downward from the funnel sinks finds its way along joints, enlarging them by solution into vertical shafts. At some levels it moves laterally along stratification planes, enlarging them into horizontal galleries. In this way a network of shafts and galleries is formed, whose development is limited downward by the water table. One by one the surface streams are diverted into this network, forming subterranean streams, which widen the galleries somewhat by abrasion. The galleries are enlarged into big caverns in the places where solution is most rapid, and the roofs of some of the higher caverns collapse, forming irregular sinks. The gradual formation of closely spaced sinks leading down into galleries leaves natural bridges, which collapse one by one, converting the rows of sinks into solution valleys (Figs. 92, 93).

Maturity. The stage of maturity is reached when subsurface drainage begins to exceed surface drainage and the greater part of the initial surface has been destroyed by the growing sinks and solution valleys, forming karst topography. Below the surface the limestone is honeycombed with caverns. One after another the streams flowing through higher galleries are tapped and led down into lower ones, as the subterranean streams gradually work down to the water table and merge with the ground water. But with the diversion of the streams from the highest caverns and galleries, the air in them becomes dry enough to permit evaporation of the water dripping from their ceilings, and the deposition of dripstone goes forward, encroaching downward level by level as successively lower caverns lose their streams.



FIGS. 91–94. An ideal cycle of erosion in soluble rocks. Dashed line = water table. Area shown, about ¼ square mile

FIG. 91. Initial stage. An area underlain by jointed limestone and drained by surface streams.

FIG. 92. Stage of youth. Formation of funnel sinks, leading part of the surface drainage underground.

FIG. 93. Stage of maturity: karst topography. Formation of caverns and additional sinks, some of which coalesce and form solution valleys. The greater part of the drainage is now underground.

FIG. 94. Stage of old age. Most of the limestone has been removed by solution, leaving only a few residuals resembling monadnocks. The drainage has worked down to the surface of the shale. From this time onward the area will be eroded still further through a fluvial cycle.

Old Age. As the last remnants of the initial surface are dissolved, and the caverns partly filled with dripstone are destroyed by successive cave-ins, maturity merges imperceptibly into old age. The surface is now a wilderness of irregular ridges and mounds separated by sinks (unroofed caverns) and solution valleys, dry with the exception of those which reach down to the top of the underlying shale. More caverns collapse, sinks continue to enlarge, and bit by bit the surface is reduced to the surface of the shale. As the last caverns are unroofed, the debris is gradually dissolved and the insoluble residue is washed away by the streams which, formerly draining the now vanished caverns, have become well developed on the surface of the shale (Fig. 94). The cycle approaches its close. The shale beneath the limestone is laid bare over much of the region, but here and there isolated monadnocks of limestone, honeycombed with cavities, remain as witnesses to the nearly complete destruction of the thick strata of which they once formed a part.

Cycles of this kind are usually confined to much more restricted areas than are fluvial cycles, because they depend on rock composition, and therefore are limited to areas of soluble strata controlled by specialized conditions. Within certain areas, such as parts of Kentucky, Indiana, France, Italy, and Yugoslavia, cycles somewhat similar to the one described are in progress, although none of them has yet attained the stage of old age.

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CHAPTER 8

LAKES AND SWAMPS

LAKES

In spite of the ceaseless effort of streams and mass-wasting to grade the land to continuous slopes, there are at work other processes that produce natural basins. Many of these basins contain lakes. As inland bodies of standing water, lakes have a large range in size, from the smallest pond up to Lake Superior, the world's largest freshwater lake, and the Caspian "Sea," the largest saline lake. The great majority of lakes lie above sealevel; some, including all coastal lagoons, are at sealevel; a few, like the Salton "Sea" and the Dead "Sea," are below sealevel. Lakes occur in all parts of the world, but, as a large number of them are direct results of glaciation, there are more lakes in high latitudes and at high altitudes than elsewhere.

Basin Making and Basin Destruction. A basin of any kind in the surface of the land is a vulnerable thing. No sooner is it formed than the agents of gradation, especially streams, start to destroy it. Valleys grow headward toward it to breach it; sediments are deposited in it to fill it up. Sooner or later every basin is effaced in these ways, and basins would be few indeed if new ones were not constantly being made. A contest between basin making and basin destruction is always in progress on all the lands.

During their existence, however, lakes are effective in modifying local climates by increasing atmospheric moisture and by cooling the air in summer and warming it in winter. Lakes of all sizes are very important as regulators of stream flow, acting as storage reservoirs and minimizing floods in lower regions to which they are tributary. They act as settling basins for stream sediments, delaying temporarily the erosion of the lands. Most of the streams tributary to Lake Erie are well loaded with mud, but the clarity of the water that spills out of the lake over Niagara Falls bears testimony to the amount of material that is dropped upon the lake floor.

ORIGIN OF LAKE BASINS

Basin-making processes are in progress in many parts of the continents, resulting in the formation of new depressions. In moist regions these basins catch water and form lakes; in many arid regions, however, they remain dry through much of the year. Nearly every major geologic process makes basins in one way or another.

Basins Formed by Crustal Movements. Such basins have existed at many times during the Earth's history, and many of them have been large. Great lake-bearing depressions caused by fracturing and dislocation of the crust (*faulting*, p. 373) are more common than those caused by bending of the rocks. The 4000-mile chain of valleys and lakes that includes the River Jordan, the Dead Sea, the upper Nile, and the African lakes, such as Tanganyika and Nyassa; was formed by the sinking down of narrow blocks of the crust between high steep walls. These great "rift valleys" contain more than 30 lakes, several of which are notably large and deep. Lake Tanganyika is 5100 feet deep, and since its surface is only 2500 feet above sealevel its bottom is 2600 feet below. Lake Baykal in north-central Asia, 5600 feet deep, the Platten See in Hungary, the Warner Lakes in Oregon, and some of the larger lakes of southern Sweden likewise owe their basins to displacements of blocks of the crust along fractures.

Slow bending of the crust may create new basins directly. It may rejuvenate a stream by increasing its gradient; on the other hand, bending against the direction of stream flow may pond the stream. Two rivers near the source of the Nile in East Africa have been ponded in this way. The Kafu River has been converted into Lake Kyoga whose main part is 140 miles long, while the Katonga River has been ponded to form Lake Victoria 260 miles long.

Some structural basins have been formed within human history in connection with earthquakes. In 1811 an earthquake shook the lower Mississippi Valley, and in western Tennessee the local sinking of the ground which accompanied it caused such changes in the surface that several new lakes came into existence.

Basins Formed by Streams. Streams form basins both by erosion and by deposition. Shallow lakes (*oxbow lakes*, p. 102) are left in meanders abandoned by the cutoff process and in abandoned temporary channels excavated during times of flood (*slough lakes*). Some of the depressions formed between natural levees and the outer margins of floodplains contain lakes such as Lake Maurepas above New Orleans.

Other basins are formed where tributaries build deposits across the valleys of larger streams. In this way Lake Pepin was formed when a 20-mile stretch of the Mississippi, 60 miles below St. Paul, Minnesota, was dammed by deposits built into it by the tributary Chippewa River. In some places this situation is reversed; the main stream aggrades its valley floor so rapidly that its tributaries, aggrading less rapidly, are ponded. The several lakes in Spokane Valley in northern Idaho and eastern Washington are of this kind.

Streams flowing over vertically jointed rocks such as columnar lava flows (p. 372) are likely to accomplish more quarrying than abrasion, and many small basins are thereby made. Many of the small basins, some of them containing lakes, on the Columbia Plateau in eastern Washington are of this kind, although a few, such as those of the lakes at Dry Falls in the Grand Coulee, are plunge basins (Fig. 41, p. 84) excavated at the bases of falls now extinct.

Basins of Marine Origin. Along many coasts, waves and longshore currents are able to build up bars offshore, converting the sheltered water back of them into lagoons (p. 237). There are long chains of these lagoons along the Atlantic coast of the United States from New York to Florida. An area newly uplifted from beneath the sea commonly contains shallow depressions inherited from the sea floor. Some of the Florida lakes are believed to be of this kind, among them Lake Apopka near Orlando.

Some lakes are cut-off arms of the sea. The Caspian "Sea" was isolated from the sea by upwarping of the land.

Solution Basins. Small lakes occur by the hundreds in sinks in extensive limestone regions such as those in Kentucky and Indiana, central Florida, Yucatan, and Yugoslavia (Chap. 7). Ordinarily the sinks that contain water are those whose bottoms lie below the water table, or whose outlets are clogged with the insoluble residue left after solution of the limestone.

Glacial Basins. Most existing lakes are the direct result of glaciation (Chap. 9). It is a striking fact that most of the lakes of North America and Europe are concentrated within areas recently glaciated. Some of these lakes occupy depressions made by glacial erosion (Fig. 95). Others occupy parts of valleys blocked by irregular glacial deposits. The basins of most of the Great Lakes in North America are believed to be former large stream valleys that were deepened by glacial erosion and somewhat altered by glacial deposition and broad warping of the crust during the glacial ages.

ALASKAN AERIAL SURVEY, U. S. NAVY.

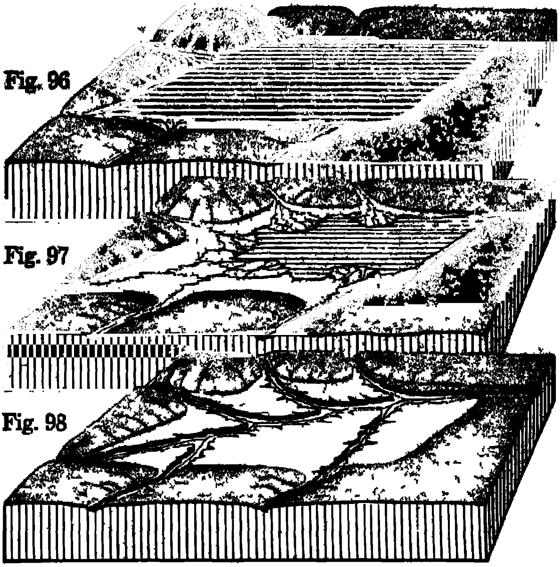
Fig. 95. Lakes occupying basins excavated by glacial erosion. Annette Island, Alaska.

Wind-formed Basins. Basins of considerable size are known to have been formed by the wind picking up and blowing away fine material such as that in beds of clay and shale (p. 202). But, since these basins are excavated only in dry regions, they rarely contain lakes except temporarily after infrequent heavy falls of rain. In some places, the depressions between sand dunes contain shallow ponds and swamps.

Volcanic Basins. Lava flows in regions where streams have already cut valleys are likely to make dams across valleys, thus forming lake basins. These barriers of lava are eroded with difficulty because they are of resistant rock. Several of the lakes surrounding the volcanic cones of Mount Hood in Oregon and Mount St. Helens in Washington are of this type.

In some volcanic regions there are lakes in the craters themselves. Crater Lake (Fig. 210) in southwestern Oregon, more than 5 miles in diameter, occupies a large caldera (p. 318). Although the lake is 2000 feet deep, it is without tributary streams, being entirely dependent for its maintenance on the fall of rain and snow directly into the basin.

Landslide Basins. Landslides (p. 60) occasionally pile up across valleys, damming streams and converting them into lakes. Several landslide lakes are known in the region of the Alps and in northeastern California. A new one was formed at the base of the Gros Ventre Range in western Wyoming by a great slide in June, 1925, which dammed the Gros Ventre River, making a lake more than 3 miles long. For some time the lake discharged by seepage through the landslide debris, but its surface rose gradually until in May, 1927, it overtopped the slide that dammed it, and the resulting flood cut a channel 300 feet wide through the dam, carried away boulders 15 to 20 feet in diameter, destroyed a village, and drowned a number of people. This flood did not destroy the lake; it merely lowered the lake surface by about 60 feet. This resulted from the great width of the dam. Because

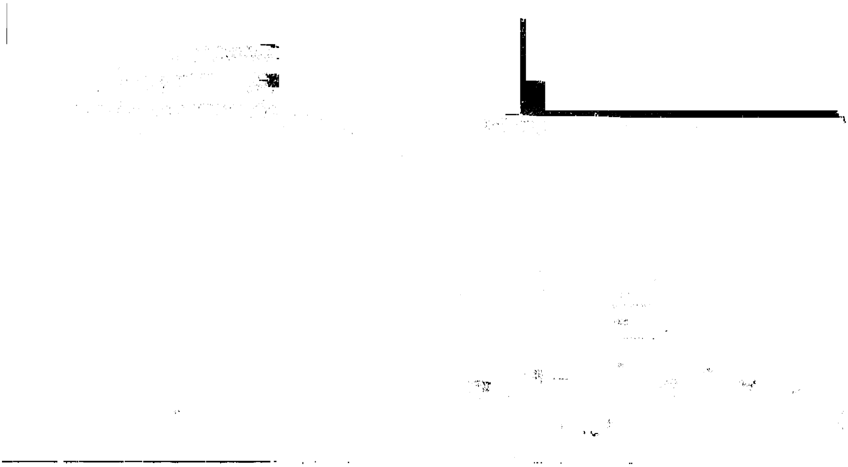


FIGS. 96–98. Stages in the history of a typical lake. Block about 5 miles long.

FIG. 96. Stream system dammed by gentle upwarp across the upper right-hand corner of the block, forming a shallow lake with outlet across the upwarp The streams begin to build deltas into the lake, and the shoreline is somewhat wave eroded.

FIG. 97. The growing deltas enlarge and coalesce, gradually filling the basin.

FIG. 98. Downcutting of the outlet eventually drains the lake and allows the streams to trench the delta deposits, leaving them as terraces.



NORTH DAKOTA AGRICULTURAL EXPERIMENT STATION.

Fig. 99. Part of the floor of the extinct glacial Lake Agassiz, seen from the air above Fargo, North Dakota. The lake deposits consist of clay and silt, and cover an area several times greater than that of Lake Superior.

of this width the energy of the floodwater that was trenching the dam was soon used up in transporting the material eroded from it. Hence downcutting ceased, and this reduced the flood to the size of a normal stream.

A somewhat similar slide occurred in a canyon near the headwaters of the Ganges, in the Himalayas, in 1893. The debris filled the canyon for 2 miles to a depth of 800 feet and ponded the river to form a lake 3 miles long. Several months later the ponded water, having slowly risen to the top of the blockade, overflowed, cutting a channel 400 feet deep through it and causing a disastrous flood downstream.

The Bonneville Dam across the Columbia River east of Portland, Oregon, occupies very nearly the site of a landslide dam that ponded the river to a depth of 100 feet in the recent past. The remnants of the slide, around which the river now detours, are still plainly visible.

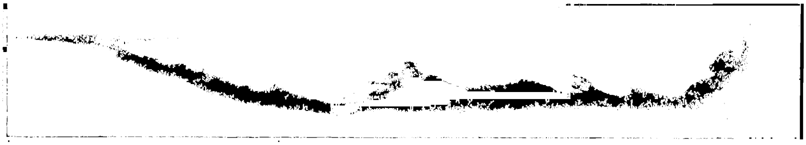
EROSION AND DEPOSITION BY LAKES

Large lakes are affected by waves of considerable size, and these waves and the currents they generate fashion the same kinds of shore features that are found on sea coasts (p. 227). In smaller lakes, however, deposition is usually predominant over erosion; the smaller and shallower the lake and the greater the load of sediment carried by the streams tributary to it, the more rapidly will it be filled up. Figures 96–98 show a small lake being progressively filled with sediment, and hen drained, leaving its deposits to be dissected by streams. Although arge lakes are less likely to be filled up, deposits may spread over their loors, forming very even surfaces (Fig. 99). When such lakes are drained, deltas are left high and dry at various points around the margins of the basins.

Lake Geneva and the Rhone. Lake Geneva, in the western Alps, typifies an early stage of the filling-up process. At present 40 miles ong, it was formerly 7 miles longer, but each year the turbid Rhone, entering the lake from the east, brings down quantities of sediment from the glaciers at its source and dumps this material into the quiet water, forming a delta that fills the upper end of the lake from side to side. The water is nearly a thousand feet deep, and the delta is correspondngly thick, but each year its bank-like front creeps farther west, and n time it will destroy the lake. While this is going on, the water that spills from the surface of the lake at its lower end is eroding its channel and lowering the lake outlet. As it passes under the many bridges of the city of Geneva, the water is clear. Nearly all the sediment from the upper Rhone has been either added to the delta or dropped upon the bottom during the slow passage of currents down the lake, so that the lake acts as a great settling basin, depriving the lower Rhone of tools with which to cut. Although downcutting of the outlet s thereby retarded, it can not be stopped. If these processes continue, the outlet will be cut lower, and the water level will correspondingly drop, until the lake will be destroyed by this process if it has not already been filled up by the encroaching delta.

Lake Erie and the Niagara. Downcutting of the rim of a large lake basin is more forcibly illustrated by Niagara Falls at the outlet of Lake Erie. The water at the brink of the falls is excavating its channel in beds of dolomite (Figs. 40, 41). At the same time the falls is retreating upstream at a rate that averages 3.4 feet per year (p. 84). Retreat of the falls is accompanied by decrease in height, because the layers of dolomite dip downward as they are traced upstream. Therefore f the falls should continue to migrate up the river to a point opposite Buffalo, Lake Erie would be largely drained. Meanwhile, the lake s being gradually filled with sediment.

Density Currents. Water containing suspended mud is slightly heavier than clear water. Where a muddy river enters a lake the coarser part of its load is deposited on the delta, and the remaining muddy water forms a *density current* (Fig. 100) that sinks to the bottom and flows along it as long as there is a slope. The current has distinct boundaries, like a cloud of dust. In fact it has been described as "a dust storm under water." Velocities of density currents vary with varying bottom gradients and amounts of mud, but rates as great as 2 miles per hour have been measured. In a long lake such a current maintains its identity through many scores of miles. Eventually the current spreads out and forms a muddy bottom layer from which the suspended sediment slowly settles to the lake floor. By the withdrawal, through a dam gate, of water from this muddy bottom layer rather than from the clearer upper water, the eventual silting up of an artificial reservoir can be postponed, and its effective life thereby prolonged.



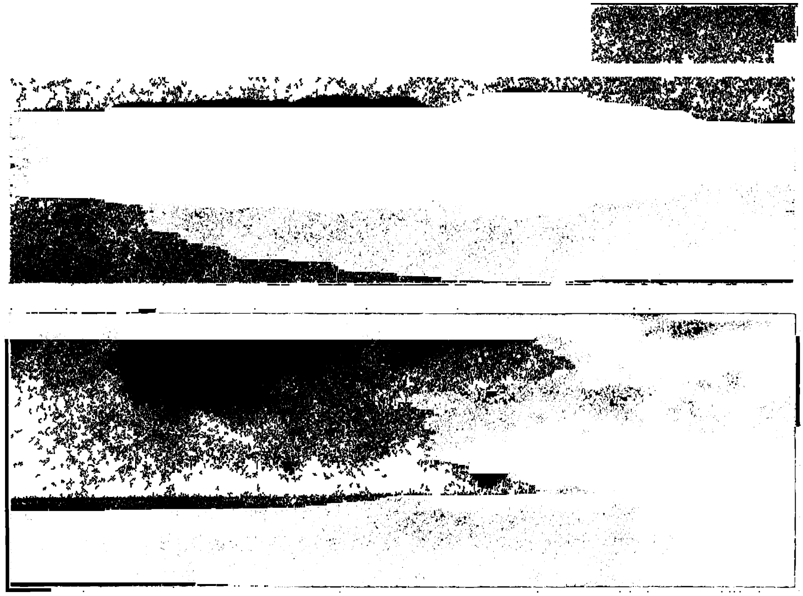
U. S. Soil Conservation Service—California Institute of Technology.

FIG. 100. Turbid underflow in a glass laboratory tank. Flow slides down delta slope (left), travels along gently sloping floor, and rises against face of dam (right). The velocity of most underflows is so small that the current does not climb far up the face of a dam

SALINE LAKES

Most of the lakes in eastern North America and western Europe have streams or springs flowing into them, and spillway outlets determined by the lowest points in the rims of their basins. Owing to the moist atmosphere evaporation takes place slowly, and the continuous inflow and outflow not only prevent the lake surface from fluctuating greatly but also keep the water fresh.

In arid and semiarid regions, however, where precipitation is slight and evaporation great, many lakes fail to rise high enough to spill out of their basins; they rise and fall seasonally, and some dry up and disappear for months at a time. The desert basins lying between the scattered ranges of Nevada, western Utah, southeastern California, and Sonora usually contain water (*playa lakes*, Figs. 101, 102; p. 113) only for short periods after infrequent rains. Wherever evaporation prevents the water from overflowing the rims of these interior basins, the substances dissolved in the water become more concentrated, and, after sufficient concentration, they are precipitated as salts of various kinds. The dissolved matter is, of course, brought into the lakes by tributary streams, which, in turn, acquired it from subsurface water (p. 128) which dissolved it from the rocks largely during the process of chemical weathering (p. 32). The salts present in a lake therefore reflect the chemical composition of the rocks within the local drainage system and are strikingly different in different lakes. Thus some lakes (*salt lakes*) when saturated precipitate sodium chloride (common salt). Others (*bitter lakes*) contain sulphates, chiefly sodium sulphate. Still others (*alkali lakes*) contain sodium and potassium carbonates. A small group (*borax lakes*) contain borax and related borate minerals. Some lakes contain combinations of these substances. Thus the precipitates from Great Salt Lake include both common salt and sodium sulphate. Deposits of all these types occur in desert basins that lie in the dry country between the Rocky Mountains and the Sierra Nevada.



C. E. Erdmann.

FIGS. 101–102. Ephemeral playa lake in a desert basin. Braun's Playa, southern Nevada.

FIG. 101. (Above) The playa filled with water, forming a shallow playa lake.

FIG. 102. (Below) View two weeks later, showing the dry lake bed after the water has evaporated. The wind is whipping up the fine lake sediments into dust clouds. The dark spots are clumps of vegetation.

In contrast, the substances dissolved in sea water are derived from the rocks of all the lands and are therefore mixed and distributed rather uniformly throughout the sea. In consequence, precipitates from sea water are distinctive. This fact aids us in distinguishing between lake precipitates and marine precipitates that have been buried in the rocks.

RECORDS OF FORMER LAKES

Lakes Bonneville and Lahontan. Great Salt Lake in Utah is the small successor to a water body, Lake Bonneville, formerly nearly as large as Lake Michigan (Fig. 103). The record of the nearly vanished lake is preserved in wave-cut cliffs, wave-built terraces (Fig. 104), bars, beaches, and deltas along the mountain sides as much as a thousand feet above the surface of the present lake. The shorelines of the former lake occur in several distinct sets, the two highest of which, traced

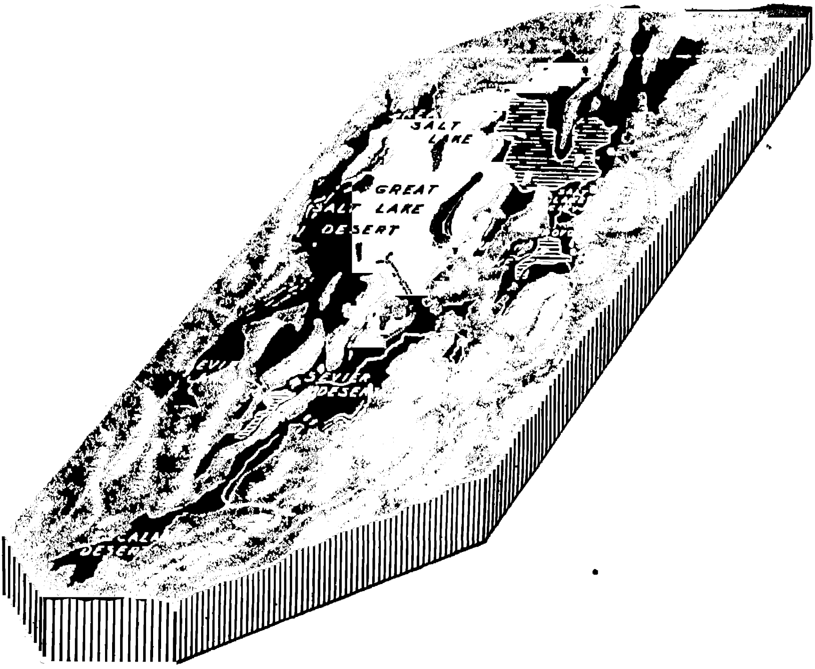
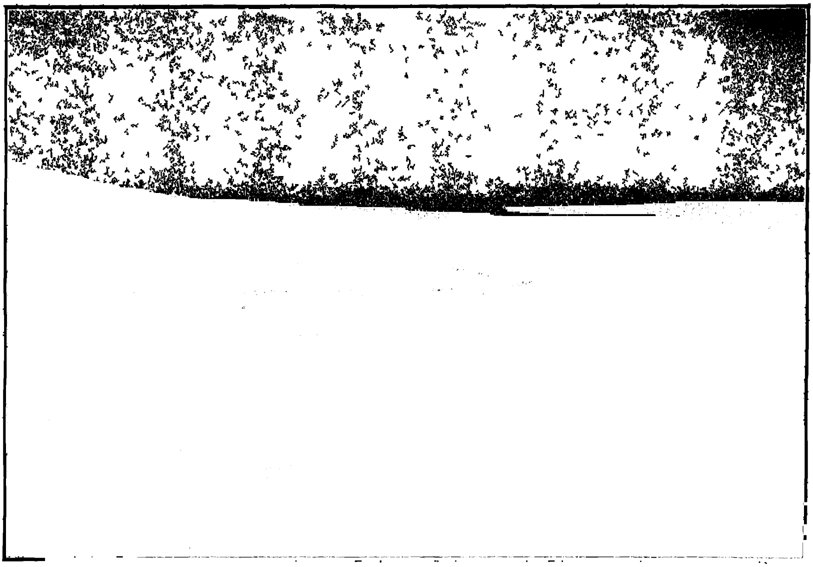


Fig. 103. Block 375 miles long and 175 miles wide, showing parts of Utah, Nevada, and Idaho. Extent of the former Lake Bonneville, when at its maximum, is shown in white. Mountain ranges formed islands and peninsulas in the lake. The outlet through which it temporarily overflowed is at the northern end of the former lake. Ruled areas are existing lakes; most of the rest of the floor of the former great lake is desert. Discontinuous abandoned shorelines marking the lake surface at various former levels are shown flanking some of the mountain ranges.

northward, merge into a broad mountain pass that leads out of the basin to the Snake River, draining to the Pacific. This pass was the outlet of the lake when its level stood relatively high. The lower sets of shorelines, unrelated to outlets, were made at times when evaporation just balanced inflow as the lake dwindled by stages. Fine sediment and salts now cover the floor of the basin.



A. E. Granger, U. S. Geological Survey.

FIG. 104. Several shorelines of the former Lake Bonneville on a spur of the Wasatch Mountains near Salt Lake City, Utah.

Similarly, Pyramid Lake and its neighbors in Nevada are the successors to another great water body, Lake Lahontan, which rivaled in size the present Lake Erie. The highest shorelines of Lake Bonneville were overridden by valley glaciers descending from the Wasatch Mountains, whereas lower shorelines, formed during a later stage of the lake, partly bury deposits made by the same group of glaciers. These facts indicate that the lake had its origin in the glacial ages (p. 189). Lake Lahontan and other former lakes in this region—at least 70 in all—also date from the glacial ages. That the climate was then more moist than now is obvious; the subsequent desiccation is clear testimony that the climate is now drier.

Salton Sea. A lake of unusual origin and history is the Salton Sea, a shallow saline lake in southeastern California near the Mexican border

(Fig. 105). It lies in the Salton basin (a part of it known as the Imperial Valley), a closed depression 85 miles long, whose lowest point is 273 feet below sealevel. In the recent geologic past the floor of this basin sank and would now be filled with sea water but for the fact that the Colorado River, turbid with silt, discharging into the Gulf of California near what is now the southern end of the Imperial Valley, had built up a combined delta and fan so large that they formed an effective dam at the head of the gulf. Like all deltas and fans this one was trenched by shifting distributary channels of the river, which discharged water sometimes into the gulf and occasionally into the isolated basin, filling it with water. When dry, the basin was kept virtually a desert under the prevailing arid climate. The last natural discharge into the basin occurred perhaps 300 to 1000 years ago. By 1900 the trapped water had dwindled to a small saline lake. By this time the fertility of the soil on the river deposits had been recognized, the basin area was beginning to be settled, and soon afterward an irrigation canal was dug from the Colorado River near Yuma to the bottom of the basin. Because the floor of the basin is well below the baselevel of the river, the gradient of the canal was greater than the gradient of the river between its mouth and the head of the canal. The

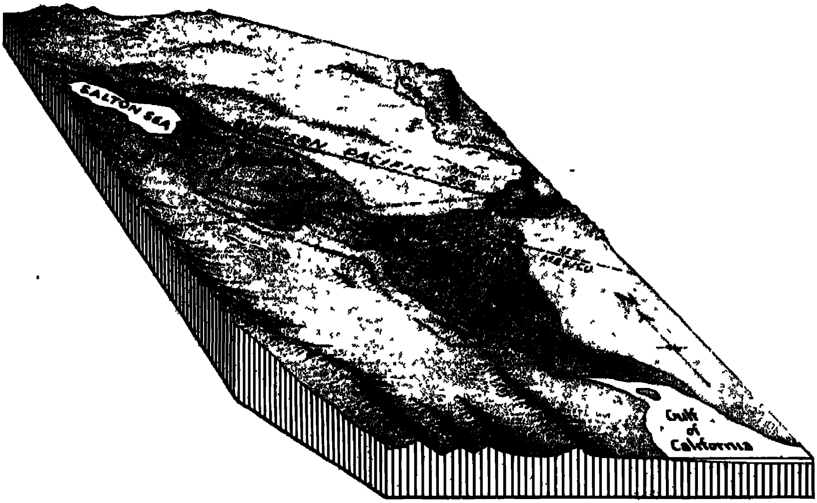


FIG. 105. The Salton basin. Block 200 miles long, showing the head of the Gulf of California, and the Salton Basin separated from it by the combined delta and fan of the Colorado River. Sealevel contour in the Salton basin is shown by thick line. Route of the flood of 1907 is indicated.

flow of water into the canal was inadequately controlled by headworks designed to be protective. In 1905, 1906, and 1907, the Colorado repeatedly rose in flood, overtopped the headworks, poured into the canal, and, following the steeper gradient, cut a great trench in some places 80 feet deep, swelled the Salton Sea to many times its former size and depth, and flooded the railroad right-of-way for 40 miles. For a long time the river resisted all efforts at permanent control, but it was finally mastered. The railroad tracks were shifted from 200 feet to 150 feet below sealevel as a precaution against possible later floods. The Salton Sea began to dwindle as soon as the abnormal supply of water was cut off, but now the waste water from irrigation keeps the level nearly constant.

Incidentally, the completion of Hoover Dam and Parker Dam, farther up the Colorado, has removed for a long time to come the flood hazard along the lower Colorado, because these dams impound part of the excess discharge in spring and release it gradually throughout the low-water period. A chain of lakes, natural or artificial, along a stream's course thus reduces floods and also furnishes traps for catching sediment. In time, however, sedimentation destroys the lakes, substituting for them a continuously graded stream profile. It is estimated that the lake created by Boulder Dam will become completely filled with sediment in 200 to 300 years, unless other settling basins are made farther upstream.

Extinct Lake Florissant. In the South Park of Colorado, in the district west of Pikes Peak, a stream is flowing through a valley whose sides are made of layers of stratified volcanic ash. These layers were deposited in a narrow lake and by related streams, and the ash came from a then active volcano not far distant. Sifting down through the lake water, this sediment protected and preserved the remains of plants and animals that lived in and around the lake. The basin gradually filled with ash and lava and was then converted into a stream valley. From the layers of ash now exposed there have been taken great quantities of fossils, including more than 1000 species of insects, 250 species of plants (including numerous trees), and many fishes and birds, all representatives of the life of the time during which the lake existed—a time that long antedated the glacial ages. The filled-up lake, known as Lake Florissant, has thus become a storehouse of great scientific value.

In adjacent regions, particularly in Idaho and Montana, similar large former lakes are recorded by their deposits. Volcanic ash is an important constituent of most of these deposits, and the fossils in them have furnished much information about the life of former times.

All the lakes now extinct have been destroyed by draining caused by erosion at their outlets, or by filling up of their basins with sediment, or by a change to more arid climate. Sooner or later one or more of these processes will destroy every lake. It is probable that the past has witnessed vast numbers of lakes, all traces of which have been destroyed.

The recognition of former lakes throws an important light upon both the climates and the basin-making processes of the past. The examples mentioned above show how it is possible to infer a former lake even in places where neither water nor closed basin exists any longer. Shorelines constitute one type of evidence; another is furnished by deltas built at the mouths of tributary streams; still another consists of the deposits spread over the floor of the lake basin, grading out and away from the deltas as bottomset beds (Fig. 99; p. 89). Ordinarily the latter are comparatively fine grained, well sorted according to grain size, and parallel bedded (Fig. 181, p. 265). Deposits in a lake that was completely filled with sediment, and thus converted into a fluvial plain (Fig. 98), grade from fine-grained sediment at the bottom to coarse-grained sediment at the top, because the gradual shoaling of the water extended the effect of the stream currents toward the center of the lake.

Swamps

Swamps are areas of saturated ground. Most swamps represent a stage intermediate between lakes or ponds and dry land. Many lakes in humid regions will in time become swamps, and many shallow basins alternately contain swamps and lakes according to the season. Swamps commonly occur in three types of regions, but these regions by no means exhaust the possibilities. (1) Coastal plains that are former sea floors slightly uplifted. Such swamps are distinguished from tidal marshes (described below) and are almost continuous along the South Atlantic and Gulf coasts of the United States, chief among them being the Dismal Swamp in Virginia and North Carolina, and the Everglades in southern Florida. It may be that some of these swamps occupy the sites of former lagoons, uplifted together with the offshore bars by which they were shut off from the sea (p. 237). (2) Floodplains and deltas with their basins formed by old channels and by natural levees. Such areas include much swamp land. (3) Broad glaciated areas such as the greater part of the Great Lakes region of the United States, eastern Canada, Ireland, and the Baltic plain of northern Germany. These regions are dotted with swamps, most of them small.

After it has been drained, some swamp land has a high agricultural value because its high content of humus makes it very fertile. Swamp lands in the United States have a combined area greater than the area of New England, and with proper draining the swamp land in some districts could be reclaimed for agriculture without creating difficulties elsewhere. In other regions, too much swamp land has already been drained (p. 92), thereby creating serious problems in excessive floods and excessive silting in streams near by. Each district has its own special problems, which should be studied thoroughly before the natural conditions of water storage and runoff are interfered with.

Tidal marshes occur only along coasts, in shallow estuaries and lagoons that are alternately submerged and laid bare by the tides. Their vegetation consists chiefly of certain grasses which grow only under such conditions. For this reason the composition of tidal-marsh deposits differs markedly from that of freshwater-swamp material.

Peat. In moist regions the shores of small lakes and small protected bays of large lakes support an abundance of aquatic vegetation such as pond lilies, water weeds, and rushes. As these plants die, their substance begins to decay in the water, largely as a result of the activities of bacteria. During the metabolism of the bacteria, toxic waste products are excreted; hence when this waste matter reaches a certain concentration in the lake water the bacteria can no longer exist, further bacterial decomposition is prevented, and the partly decomposed matter is preserved. This matter is brownish or blackish, has a high carbon content, and is known as *peat*.

As peat is built up near the shore, newer generations of aquatic plants advance toward the center of the lake, and other types of vegetation such as mosses encroach over the peaty area that was formerly water. In this way the lake, surrounded by concentric belts of different kinds of plants, gradually decreases in size until it is obliterated, and a swamp or bog floored with a thick accumulation of peat takes its place (Fig. 106).

In many countries, especially in Europe, peat is cut from the bogs, dried, and used as a domestic fuel. When dried, peat is a dark-brown or blackish fibrous substance, very light in weight. In America, peat has hitherto been little used because of an abundance of coal and wood. Nevertheless the peat resources in swamp lands within the United States are enormous and constitute a valuable potential source of fuel for the future.

Peat is of special interest to the geologist in that it represents the first stage in the transformation of vegetable matter into coal (Chap. 21). A complete gradation can be traced from peat through lignite, bituminous ("soft") coal, and anthracite ("hard") coal. If the peat bogs of today were left untouched, some of them would in the course of time be covered with sediment, the remaining necessary changes would take place, and the result would be the formation of coal interbedded with other sedimentary rocks.

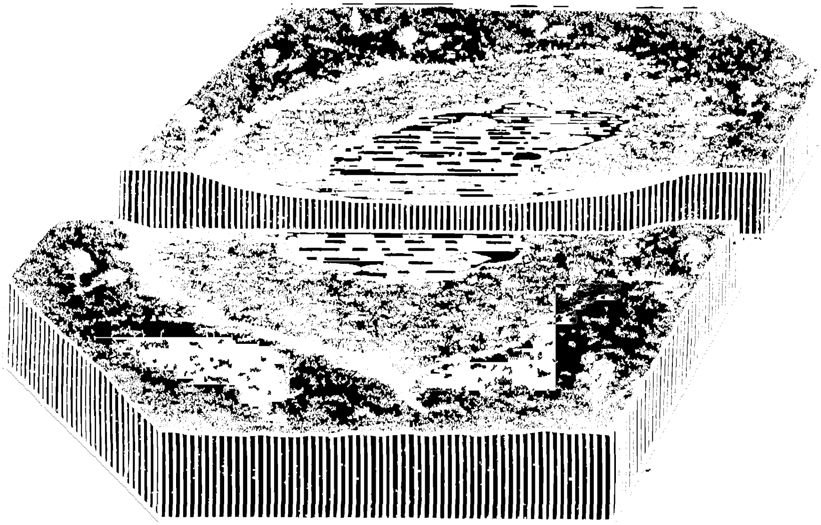


FIG. 106. Destruction of a small lake by filling with peat. Vegetation growing from the shores toward the center of the lake is gradually filling it up. The accumulating peat (black) is fringed by aquatic vegetation, which in turn is being encroached upon by semi-aquatic plants, mosses, and bushes.

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CHAPTER 9

GLACIERS AND GLACIATION

Geologic Role of Glaciers. To anyone who has observed with a critical eye the Alps, the Sierra Nevada, or the higher ranges of the Rocky Mountains, it is obvious that their pinnacled peaks and trough-like valleys are unlike those sculptured by streams in lower mountain ranges. The trained observer would at once recognize these alpine forms as the work of glaciers, even though the glaciers themselves had disappeared. A study of the character and behavior of these striking and impressive sculptors of the land helps us to understand not only the way in which they operate, but also the basis for the universal opinion that glaciers have recently extended over vast areas now free of ice, altering the landscape, redistributing the mantle, and conspicuously changing the level of the sea. An inquiry into the geologic effects of glaciers also enables us to learn much about the climates of the past.

Types of Glaciers. We can readily distinguish two chief kinds of glaciers, on the basis of their form. Most numerous and most familiar are *valley glaciers* (Figs. 108, 133) which generally flow downward from high snowfields through mountain valleys. They are long and narrow, like swollen but sluggish streams. Less familiar, because they are now mainly confined to high latitudes, are *ice sheets* (small ones are often called *ice caps*) (Figs. 107, 111). These are very broad (in some cases nearly circular) masses of ice lying on surfaces of low relief. Some small ice caps are hardly a mile in diameter; in contrast the vast ice sheet on the Antarctic Continent has an area of about 5 million square miles.

STRUCTURE, FLOW, AND WASTAGE OF GLACIERS

Distribution of Snow. Snow is controlled by climate. In the tropics it falls only on the highest mountains and plateaus. In middle latitudes it falls on lowlands as well, but disappears in summer. Over many parts of the cold polar regions it covers wide areas and remains from year to year. *Snowfields,* areas of perennial snow, in which summer wastage (melting and evaporation) fails to remove winter snowfall, are therefore found only in high latitudes or at high altitudes. The general lower limit of perennial snow in any region is the *regional snowline*, and its general position is determined by climate. It rises from near sea-level in Polar regions to altitudes of more than 20,000 feet in dry low-latitude regions such as the northern Andes and Tibet. Local differences of precipitation and wastage also exert strong effects. If the drier side of a mountain happens also to be the side more exposed to sunshine, the snow limit is much higher there than on the moister, more shaded side.

Controlled by ever-changing combinations of climatic factors, the regional snowline in any one place shifts slightly from year to year, and over periods of hundreds or thousands of years the shifts, in some mountain districts, amount to hundreds and even thousands of feet. Rise or descent of the regional snowline is therefore a useful register of slowly changing climate in the region where it occurs.

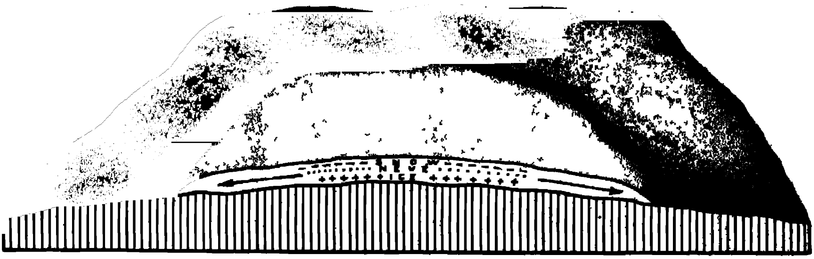


FIG. 107. Half of a small, nearly circular glacier of the ice-cap type built by snowfall on a smooth plateau-like surface. The mass of snow, névé, and ice has become thick enough to acquire motion, the general directions of which are shown by arrows. The ice is free to spread outward in all directions.

Conversion of Snow into Ice. When the snowline in any region descends, the glaciers in that region expand, and when the snowline rises, the glaciers shrink. This is because the glaciers result from the piling up and compacting of snow: the more snow, the larger or more active the glaciers. Under the influence of moisture and the gentle pressure exercised by the weight of new snowfalls, each snowflake is consolidated into a little ball of ice, and the resulting mass (*névé*, pronounced "nayvay") has the granular texture that we find in the last lingering remnants of any snowdrift. If more snowfalls continue to add their weight to the mass, part of the air between the granules is squeezed out, and the névé gradually becomes compact ice. The ice nevertheless betrays its snowflake origin by the fact that it is distinctly layered, each layer representing a single snowfall. The planes between some of the layers are sprinkled with films of blown dust, making the stratification even more distinct.

Conversion of Snowfield into Glacier. When the whole mass, consisting of compact ice at the base, névé in the middle, and fresh snow at the top, has been built up to a critical thickness (usually 100 to 150 feet) that depends partly on the steepness of the slope on which it is

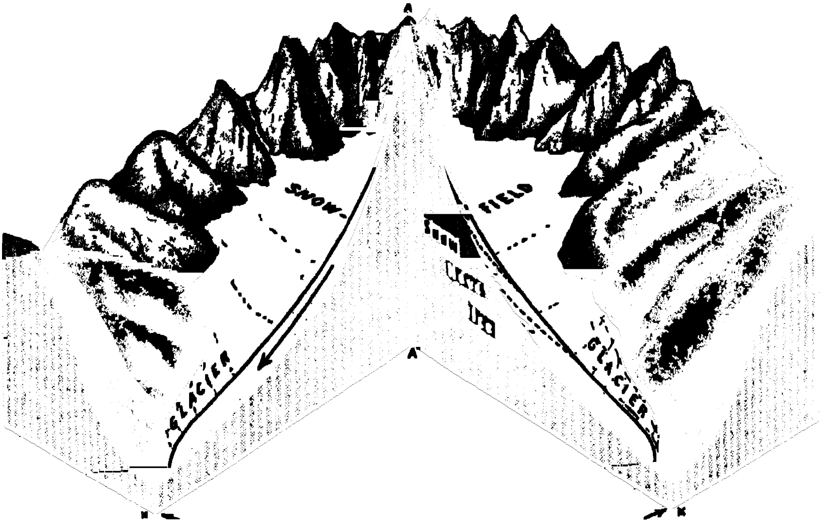


FIG 108 Snowfield with small valley glacier, in a broad niche or basin near the crest of a mountain range The ice has become thick enough to flow, but it is free to spread in one direction only—down the valley to which the niche is tributary (Compare Fig 133) Hence it takes on a tongue-like form

In order to reveal a section of the snow and ice, the block has been cut and pulled apart around a vertical axis *AA* The point *x* is the same on both halves of the block. The arrow indicates the general direction of flow of the ice.

resting and partly on temperature, it begins to creep gently downhill, flowing outward and downward (Figs. 107, 108). The creeping or flowing mass is a *glacier*, and, as long as it continues to receive an adequate supply of snow, it will continue to flow.

The snowfield is not distinctly separated from the flowing (glacier) ice. Even the ice in the snowfield is in motion, because the particles of fallen snow are imperceptibly transferred downward and outward into what is definitely the glacier.

Character and Flow of Glacier Ice. The flowing movement of the glacier is somewhat like that of a viscous fluid like asphalt. Glacier ice,

BRADFORD WASHBURN.

Fig. 109. Formation of an iceberg by calving at the terminus of Miles Glacier, near Cordova, Alaska, in a lake. A few seconds before the photograph was taken a large block of ice slid off the 150-foot cliff and started a series of waves. The intricately crevassed surface of the glacier has been melted into fantastic shapes.

however, consists of grains of solid crystalline ice each a fraction of an inch (or more) in diameter. As weight is piled upon it, the ice in the basal part of the glacier becomes plastic and yields. It is therefore the lower part of a glacier that flows; the part near the upper surface, having little weight upon it, is nearly or wholly rigid. The rigidity of the upper ice is demonstrated by the cracks and *crevasses* that commonly occur in it. These are superficial features and do not descend far into the flowing ice beneath.

Measurements made on the upper surfaces of valley glaciers indicate that their flow differs markedly from place to place within the glacier. The velocity of flow along the central axis may be many times that along the lateral margins. This is chiefly because friction is less at the center than at the sides as in a stream

Wastage. Wastage is the direct or indirect conversion of ice into water or water vapor. It takes place by evaporation, melting, and calving. Calving (the breaking off of pieces of ice, usually as icebergs) occurs commonly only at the glacier's terminus and is general only in high latitudes (Fig. 109). Evaporation and melting, however, affect the entire surface from snowfield to terminus and therefore reduce the discharge of ice. Wastage in the snowfield reduces the amount of ice transferred to the glacier proper, and wastage of the glacier's surface takes its toll throughout its length, especially in the terminal zone, where temperatures are higher than at the glacier's source.

Meltwater. The most conspicuous product of wastage is meltwater. It forms streams and pools on the upper surfaces of many glaciers in their terminal zones; it concentrates into streams along the lateral margins of glaciers in valleys; it descends through crevasses and tunnels to the bottom of the glacier. Most glaciers have swift streams of meltwater flowing away from their lower ends.

REGIMEN OF GLACIERS

Normal Regimen of a Glacier. The regimen of a mass of snow or névé that is too thin to flow is controlled by the interplay of two factors: *nourishment* by snowfall, which tends to increase the mass, and *wastage,* which tends to diminish it. A preponderance of snowfall over wastage through a long enough time will increase the mass until it becomes thick enough to flow. At this moment a third factor is introduced into the regimen: the transfer of ice from one part of the glacier to another by flow. The greater the excess of nourishment over wastage, the more rapid the rate of flow.

Every glacier therefore has in its higher part an area of dominant nourishment and in its lower part an area of dominant wastage. The ice flows from its higher part toward its lower part, thus making good the losses sustained through wastage. This is true regardless of the shape, form, or size of the glacier, as can be visualized by inspection of Figs. 107 and 108. If both nourishment and wastage remained unchanged from year to year, the downstream transfer of ice by flow (that is, the discharge, as in a stream) would be uniform, and the regimen of the glacier would be in equilibrium. Actually, however, equilibrium is seldom attained because both snowfall and wastage vary from year to year.

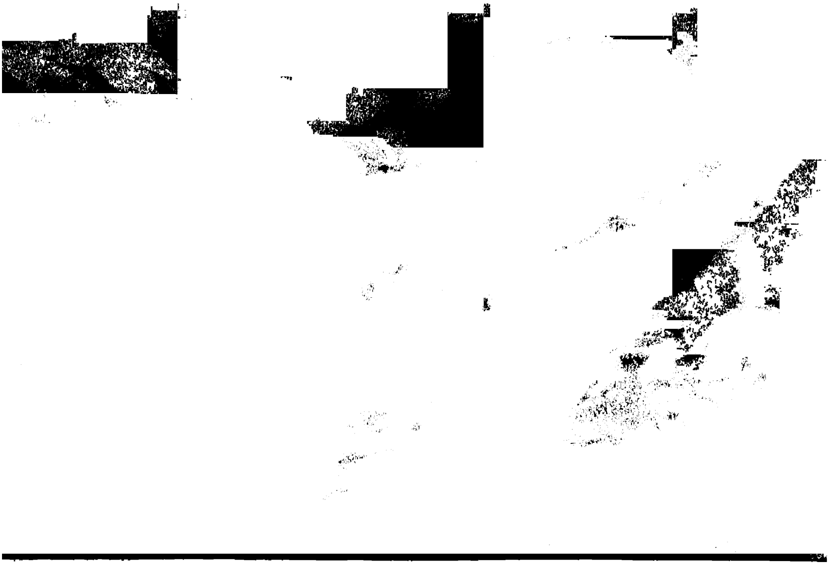
The regimen described is much like that of a stream of water that drains a shallow lake, flows down a mountain valley into a desert, and there disappears by evaporation. In such a stream the three factors of rainfall, stream flow, and evaporation are so closely interrelated that even a slight change in one factor affects both the others. The larger the area of nourishment and the greater the snowfall upon it, the greater the discharge of ice through the glacier. Thus a narrow thin glacier must flow more rapidly than a wide thick glacier in order to keep its discharge adjusted to the amount of snowfall in a given area of nourishment.

For these reasons, no two glaciers are likely to flow at the same rate. Furthermore, the rate of flow of any one glacier decreases with decrease in temperature (some glaciers are nearly at 0°C.; others are much colder) and other conditions, not only from season to season but also in some instances from hour to hour. Measurements of flow range from a small fraction of an inch per 24-hour period in certain Antarctic glaciers (probably the coldest in the world) to many tens of feet.

Fluctuations of Glaciers and their Causes. Individual glaciers watched from year to year are found to become either longer and thicker over a period of years or, as happens to be true of most glaciers at present, shorter and thinner. These fluctuations are the result of the changes in nourishment (dependent chiefly on amount of snowfall) and in wastage (dependent chiefly on summer temperatures). As both snowfall and temperature are basic elements in determining climate, it is apparent that similar fluctuation of a large group of glaciers over a long period of years indicates a change in the climate.

It is much easier to measure and record changes in the position of a glacier's terminus than changes in its thickness. Hence measured fluctuations in the terminus are commonly used as general indicators of climatic change. Long records have been kept on some of the more readily accessible glaciers in the Alps and in western North America. Nearly all the glaciers are shrinking (Fig. 110). For example, on Mount Rainier in Washington, Nisqually Glacier, with a present length of about 4 miles, was shortened 4131 feet from 1857 to 1944. The average annual recession of this glacier's terminus was thus more than 47 feet during the 86-year period, though the figures show that after 1930 the rate of recession increased to nearly four times the average rate prior to 1892.

The neighboring Paradise Glacier on Mount Rainier, and a number of other glaciers in western North America, have shrunk to such an extent that their terminal parts have become stagnant, lacking further flow. This condition may result from such extreme thinning by wastage



BRADFORD WASHBURN.

Fig. 110. Nunatak Glacier, Alaska, photographed in 1934, showing position of its terminus. The approximate position of its terminus in 1909 is shown for comparison. The glacier was shortened by 6 miles during the 25-year interval, chiefly by breaking off piecemeal into Nunatak Fiord. Small icebergs are visible in the view. Contrast the ice-smoothed lower slopes of the valleys with the frost-wedged upper slopes. Sinuous medial moraine records flow through a sinuous valley.

that the ice no longer has sufficient thickness to make the basal part flow, and the upper more or less rigid zone of fractures and crevasses extends right down to the base of the ice. When crevasses cut the ice from top to base, masses of ice may thereby become isolated from the main body of the glacier. In two glaciers on Mount Shasta, California, the gaps between large isolated ice masses and the termini of the continuous glaciers were found in 1936 to be 1 mile and 1½ miles respectively. The initial severance of one of them is believed to have occurred in 1920.

The shrinkage of glaciers throughout the world since the middle of the nineteenth century is a sound basis for the belief that during the past hundred years or so the climates have grown slightly warmer.

In contrast to the general shrinkage which most glaciers are now undergoing, marked expansions of a few glaciers occasionally occur.

Thus in September, 1899, severe earthquakes affected the district of Yakutat Bay in Alaska, where there are many glaciers in a nearly stagnant condition. By 1906 most of the glaciers were expanding rapidly, but later they returned to their former sluggish flow. The sudden expansion appears to have been the result of enormous quantities of additional snow avalanched by the earthquakes from the surrounding cliffs into the snowfields that fed the glaciers. The augmented flow was analogous to a sudden flood in a stream system, brought on by a cloudburst. The 7-year lag, however, is a result of the slight mobility of glacier ice as compared with water.

The recent record of the Black Rapids Glacier, in another part of Alaska, is even more striking. During a period of 5 months in 1936-1937, flow was so rapid that, in spite of wastage, the terminus advanced 3 miles—a daily average of 115 feet, possibly the highest rate on record for any glacier. Exceptionally great precipitation occurred in this district during the period 1929–1932. It is thought that, after a lag of several years, this excess was making itself felt at the terminus. The glacier is small in proportion to the snowfield from which it is nourished; this helps to explain its unusually rapid flow.

Genetic Relations of the Glacier Types

Valley Glaciers. It has been stated (p. 158) that the commonest type of glacier is the valley glacier. Flowing from mountain snowfields, these glaciers take the lowest routes available, following the valleys that had been cut by streams before the glaciers came into existence (Figs. 109, 110, 124, 133). In the Alps there are more than 1200 valley glaciers, the greatest of which is nearly 10 miles long. Valley glaciers lie high in the other ranges of Eurasia—the Pyrenees, the Carpathians, the high mountains of Norway, the Caucasus, and the Himalaya, to name only a few. The lofty mountain valleys of the Andes and of New Zealand carry many large glaciers, and in the mountains of the Alaskan coast thousands of valley glaciers, favored by great precipitation and cool temperature, reach down to or near the level of the sea. The west branch of Hubbard Glacier, Alaska, 75 miles long, is the longest valley glacier yet reported. Southward through British Columbia, Washington, and Oregon, the regional snowline rises, and glaciers become less numerous. Within the United States the largest glaciers occur on the high volcanic peaks of the Cascade Mountains, of which the most important are Mount Baker, Mount Rainier, and Mount Adams in Washington, Mount Hood in Oregon, and Mount Shasta in northern California. Some are present also in the Rocky Mountains.

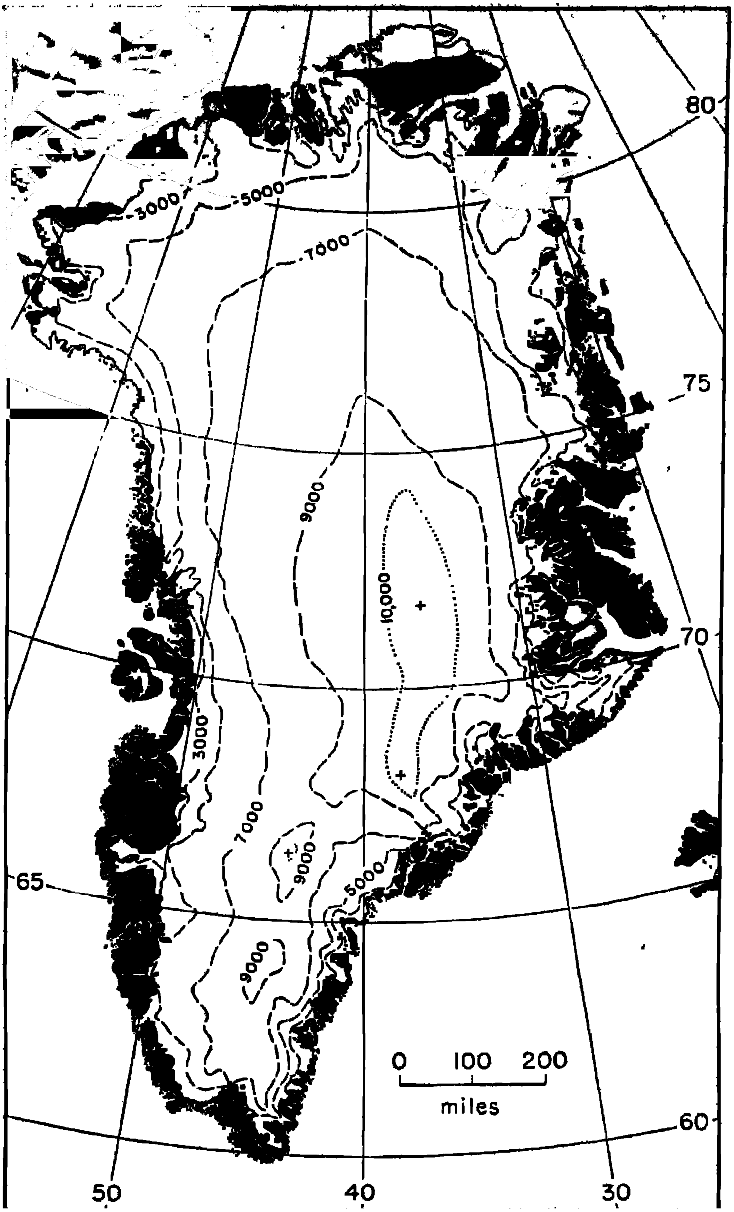
In the Alps there are in addition to the valley glaciers many fields of snow, névé, and ice, lacking perceptible motion, and too small in size to give rise to actual glaciers. These small fields persist in regions that are not quite cold enough or moist enough to nourish full-sized valley glaciers. This is true even of high ranges of the Rocky Mountains in Montana, Wyoming, and Colorado.

Piedmont Glaciers. Some valley glaciers emerge from mountain ranges and spread out upon plains, coalescing to form a continuous thick sheet of ice. The greatest of these *piedmont glaciers* are Malaspina Glacier (with an area of 1500 square miles) and Bering Glacier, both in Alaska. They are intermediate in character between valley glaciers and ice sheets.

Ice Sheets. Small ice sheets (or ice caps) are numerous on high-latitude plateaus, where they have been formed entirely by the accumulation of fallen snow on the plateaus themselves. In addition to these, much larger ice sheets exist. One of these is the mighty Greenland Ice Sheet, a continuous mantle of ice that is not confined by valley walls, but rises above valleys and mountains alike to an altitude of more than 10,000 feet. This great glacial blanket has an area of more than 637,000 square miles and is thick enough (6200 feet at one point near the center) to cover high mountains and plateaus, burying them completely. Its form, though imperfectly known, appears to consist of two or more broad low domes (Fig. 111), with almost imperceptible slopes. From the central part of the mass the ice slowly creeps toward the coasts, flowing through valleys that transect high coastal mountain ranges. Most of these outlet tongues of ice, which resemble valley glaciers, end in the sea where they discharge great pieces of ice that float away as icebergs.

The Antarctic Continent is covered by a similar but much larger ice sheet having an area of more than 5 million square miles. Partly through outlet tongues and partly as broad unbroken sheets its discharge reaches the sea and forms bergs.

The surfaces of these two great glaciers are broad monotonous wastes, barren of life. The only visible movement upon them is the movement of wind-driven snow and particles of ice. Other very large ice sheets formerly overran parts of North America and Eurasia, as detailed elsewhere in this chapter.



Modified from Flint, Glacial Geology and the Pleistocene Epoch.

FIG. 111. Greenland, showing the Ice Sheet and other glaciers in white; ice-free areas are dark. Surface form of the Ice Sheet is indicated by generalized contours (altitudes in feet above sealevel). Crosses mark high points on the ice-sheet surface.

It is very improbable that the present great ice sheets and those of former times were built up, like their relatives the small plateau ice caps, entirely by snowfall on the areas they covered when at their greatest extent. They are believed to have resulted from the gradual invasion of their territory by valley glaciers and ice caps flowing from higher places, the invading glaciers having coalesced to form a single glacial mass. Further snowfall on the coalesced mass built it up so thick that it partly or entirely buried the mountains from which the invading glaciers had descended. According to this view a large ice sheet is somewhat like a large lake formed by the mingling of waters from many streams that flow into it from surrounding mountains. The lake level rises, submerging an ever-increasing expanse of territory. In any case there are no sharp distinctions between valley glaciers, piedmont glaciers, and ice sheets. The three types grade into each other.

GEOLOGIC WORK OF GLACIERS

By *glaciation* is meant the alteration of the surface in consequence of glacier ice passing over it. It involves both erosion and deposition, each of which is described separately below.

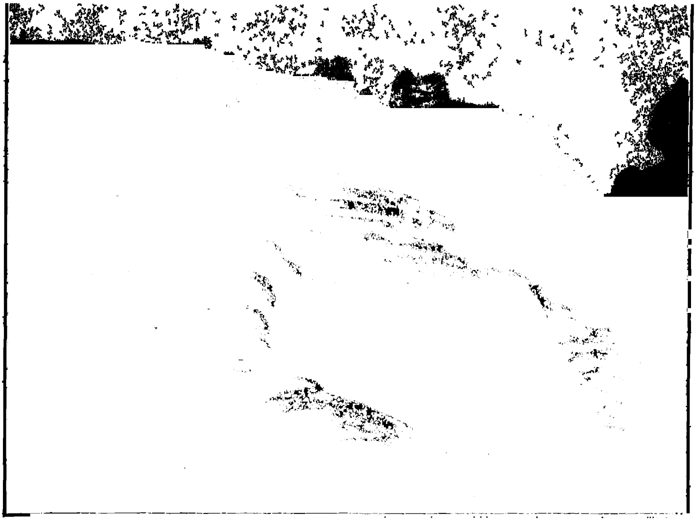
GLACIAL EROSION

Plucking and Abrasion. It has been said that a glacier is at once a plow, a file, and a sled. As a plow it churns up and moves mantle and pieces of the bedrock; as a file it rasps away the firm rock in its path, smoothly abrading it; as a sled it carries away the plowed-up and filed-off debris, plus whatever additional rock fragments may have fallen from mountain slopes down on to its upper surface.

The chief process in glacial erosion is *plucking*, which can be thought of as including all the ways in which debris is picked up and incorporated into the moving ice. Plucking is the plowing process mentioned above and is broadly the equivalent of hydraulic action in streams of water (p. 76). The ice in the base of the glacier drags forward loose rock fragments on the ground underneath it or freezes around pieces of rock, incorporating them into the base of the ice. The pressure caused by the weight of the thick moving glacier breaks off and quarries out blocks of bedrock, especially from surfaces that are unsupported on their lee sides. Such plucking of bedrock is aided by frost wedging (Fig. 14, p. 35). Meltwater percolates into joints and cracks in the bedrock, freezes, expands, and thereby pries out and shoves forward blocks of rock.

Glacial erosion also includes *abrasion*, the filing process. Like a stream of water, a glacier cannot abrade without tools. But its under surface is studded with rock fragments of many sizes, and with these as teeth the slowly moving ice bites slowly into the bedrock underneath, making long scratches and grooves (*glacial striations* or *striae*).

The effects of drastic plucking and abrasion on the surface of the bedrock are conspicuous. In Fig. 112, a glacier has plucked large blocks from the bedrock. It has filed and polished the parts of the rock that it could not remove by plucking, as shown by the striations and polish on its surface. From the distribution of the striations and the quarried rock surfaces in the picture it is easy to infer that the former ice was moving from left to right. Figure 113 shows striations in greater detail.

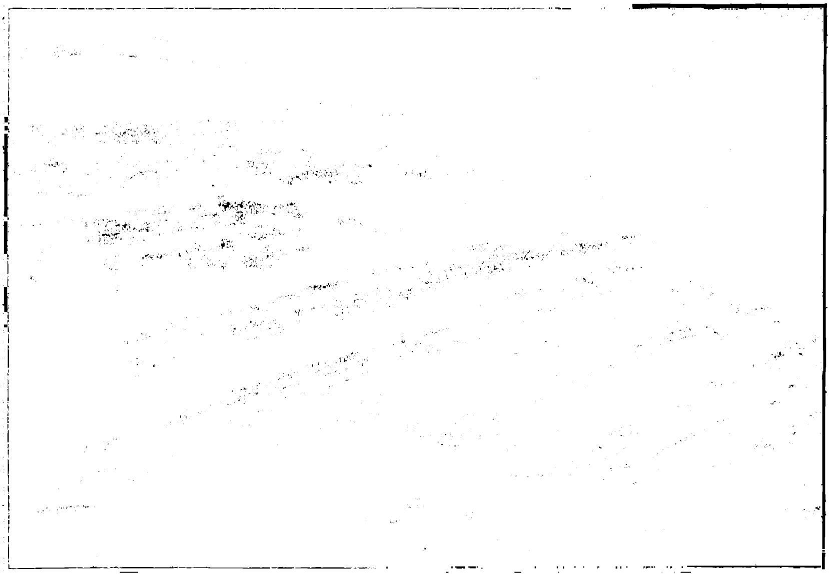


G. S. Young.

FIG. 112. Grooved and polished surface of bedrock, Middle Fork of Kings River Valley, California. The former ice (a valley glacier) flowed from left to right. The abraded side (left) differs distinctly from the quarried side (right).

While the bedrock surface is being abraded, the stones in the base of the glacier are themselves worn, usually unevenly, because from time to time they are forced to turn in their icy matrix. An ideal glaciated stone (Fig. 142, *A*) therefore has several facets, some of them bearing striations, the facets meeting each other along somewhat rounded edges.

In general, glacial erosion is accomplished more through quarrying than through abrasion, because it takes less work per unit volume to quarry out most kinds of rock than to wear rock down by rubbing.



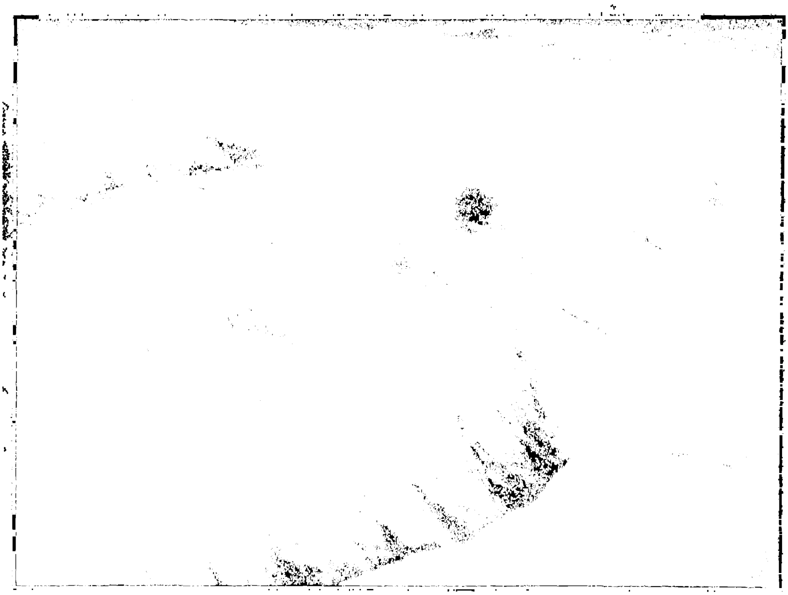
K. S. Brown, U. S. Bureau of Reclamation.

FIG. 113. Glaciated surface of bedrock (basalt), showing grooves and striations. A few ice-transported stones lie on the surface. Disintegration by weathering since the glacier wasted away is visible in the foreground. Near Grand Coulee Dam, Washington.

GLACIATED VALLEYS

Glaciers do not excavate valleys of their own: they remodel valleys already formed. Both the valley glaciers that creep downward from the snowfields near the crest of a mountain range, and the long tongues of ice that are the advance guard of an ice sheet flowing over a mountainous region, follow the valleys that had been excavated by streams before glaciation began. The glaciation of a mountain valley begins at its head.

Valley Heads: Cirques. Valleys that are or have been occupied by valley glaciers have steep blunt heads (Figs. 114, 134). These striking valley heads, known as *cirques* (pronounced "serks"), probably are started by the gradual excavation of niches that are occupied from year to year by snowbanks at the heads of mountain valleys. On summer days the melting snow forms water which permeates the rock beneath the snowbanks, especially beneath their margins. At night the temperature drops, and the water freezes and expands, breaking up the surface rock. The smaller rock particles are washed away by meltwater during daytime thaws. All this activity very slowly enlarges the depressions beneath the snowbanks. The depressions therefore gradually take on the form of the snowbanks, becoming broader and flatter floored and deeper (Fig. 115). If a snowbank increases



Sydney Bonnick for Alexander Forbes.

FIG. 114. View near Eclipse Harbor, northern Labrador, across the Torngat Mountains, showing bowl-like cirques at the heads of glaciated valleys. Some of the cirques still contain snowbanks.

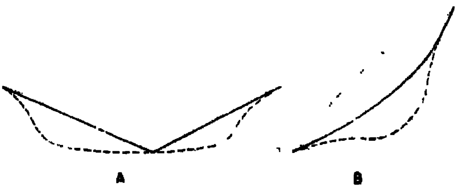


FIG. 115. Alteration of the cross profile (*A*) and long profile (*B*) of a mountain niche, to form a cirque. Continuous line = initial profile of the rock surface. Dotted line = surface of the snowbank. Broken line = profile of the rock surface after alteration by snowbank erosion.

enough to take on motion and become a small glacier, the resulting plucking and abrasion help the meltwater process to enlarge the depression into a full-fledged cirque. Erosion is so intense in the floor of a cirque that many cirque floors are excavated to form rock basins. When, after warming of the climate has melted away the ice and snow, these basins are exposed, they fill with water and form small, nearly circular lakes. Cirques are not made by ice sheets, because ice sheets overtop the rock surfaces in the regions where they occur.

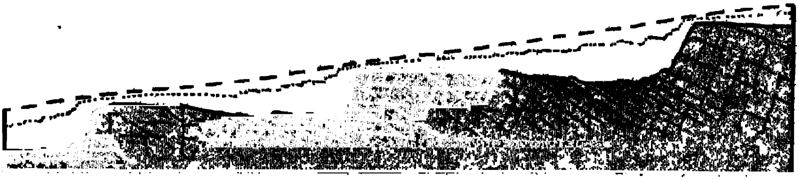
F. E. Matthes, U. S. Geological Survey.

FIG. 116. Valley abandoned by a valley glacier after the glacier had beveled its spurs, smoothed its sides, deepened it, and given it a U-profile. Tenaya Canyon, Yosemite Valley, California.

In the process by which cirque excavation is begun the reader will recognize *frost wedging* (Fig. 14). This process goes on not only beneath glaciers in regions warm enough to permit subglacial melting, but also on exposed bedrock surfaces above the glaciers. Alpine peaks owe much of the detail of their jagged form to this process.

Other Features of Glaciated Valleys. Glacier ice, being far less mobile than water, does not conform so readily to the bends and curves (Fig. 56) imparted to the valleys by the streams that originally excavated them. For this reason it grinds persistently against the spurs that project alternately into the valleys, snubbing their ends and gradually beveling them into facets. The facets grow larger as the spurs grow shorter, until at length the spurs are worn away entirely, leaving wide and nearly straight U-troughs through which glaciers can flow with minimum effort (Figs. 116, 119).

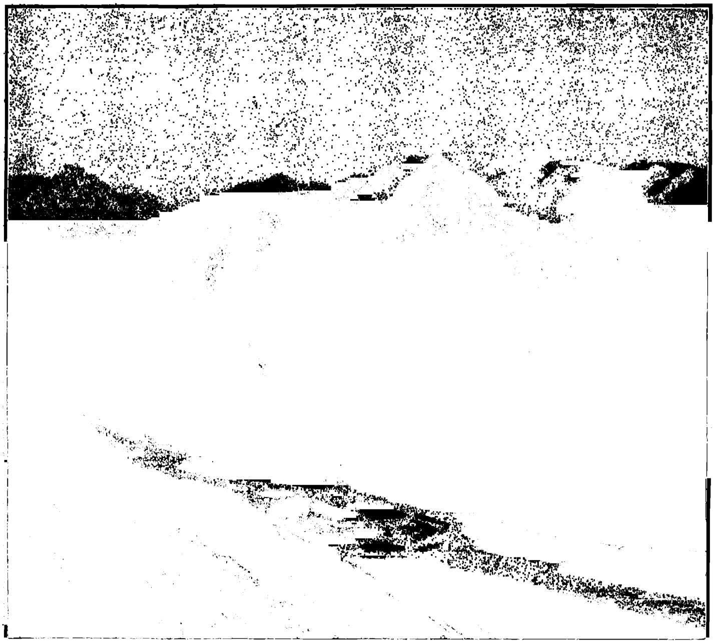
The long profiles of some valleys are greatly altered by glaciation. More or less smooth stream-valley profiles are remodeled by valley glaciers into short treads alternating with steep risers, resembling great flights of stairs (Fig. 117). In some places the outer parts of the treads are cut from rock comparatively free from joints and fissures, whereas the bases of the risers are cut into thoroughly jointed and fissured rock. This seems to indicate that some of the stairs are made by quarrying in the jointed rocks and by abrasion in the places where the joints are scarcer. In the latter places the rock surfaces are commonly polished, grooved, or striated. Some valleys that have been occupied by vigorous glaciers contain, in addition to stairs, basins excavated in the bedrock. Many of them now contain lakes. Such basins are common in many glaciated valleys and are rare in nonglaciated valleys.



After F. E. Matthes.

FIG. 117. Long profile and section of part of a glaciated valley showing smoothed surfaces of glacial abrasion alternating with steep faces caused by glacial quarrying, imparting to the valley a step-like profile. In this valley the steps are shown to be controlled by the unequal distribution of joints and other planes of weakness that cut the bedrock, although other factors may affect the making of such steps. Dashed line indicates valley profile prior to glaciation; dotted line indicates profile during an early phase of glaciation.

Hanging Tributaries. The grinding away of the spurs in a main valley affects the tributary valleys, whose mouths are snubbed back at a rate faster than their own streams (or their thin and relatively weak glaciers) can deepen them. The tributary valleys can no longer enter the main valley at grade but are left hanging above it, so that after the ice has disappeared their streams flow out to the edges of cliffs down which they must cascade in order to join the main. These *hanging tributary valleys* (Fig. 118) are characteristic of regions sculptured by glaciers, although somewhat similar hanging tributaries originate in other ways (p. 492). Most hanging tributaries result from deepening of their main valleys by valley glaciers. However, some have been made by ice sheets that swamped entire valley systems. The ice sheets deepened the valleys lying parallel with ice motion, more than the transverse valleys, which tended to fill with glacial deposits.



Geological Survey of Canada.

FIG. 118. Valley of Taku River, British Columbia. Widening and deepening by a valley glacier have converted it into an open trough and have left the tributary valley (right) hanging above the floor of the main valley, forcing the tributary stream to cascade down a cliff.

The hanging tributaries of the Finger Lakes valleys in central New York are believed to have resulted from ice-sheet glaciation. If the cut-away parts of their profiles are restored by projecting them into the main valley (Fig. 119) they are commonly found to meet above the floor of the latter. This proves that the main valley has been not only widened by the ice but deepened as well. The Yosemite Valley in California, originally cut by a stream, was later deepened as much as 1500 feet by glaciers that occupied it for a considerable time.

Fiords. In middle latitudes the features that show a valley has been glaciated disappear as the valley is traced downward, and the lower limit of glaciation, the place where the glacier terminated, can therefore be approximated. In high latitudes, however, most valleys, even along the coasts, have been glaciated throughout their lengths, and are now in part *fiords*. A fiord is a segment of a glaciated trough that is partly filled by an arm of the sea (Fig. 110). It is a characteristic feature of the coasts of British Columbia and Alaska, Labrador and Greenland, Norway, Chile, and parts of New Zealand. The Franz Josef Fiord in East Greenland is 100 miles long and 8 miles wide near its mouth, and in places its floor lies more than half a mile below sealevel. Messier Channel in Patagonia, the deepest known fiord, reaches a depth of 4250 feet Such great depths may be due in part to regional submergence of the coasts in addition to glacial excavation. However, some fiord floors include bedrock basins which, if brought above sealevel, would still contain lakes hundreds of feet in depth. Such features must have been excavated by powerful glaciation.

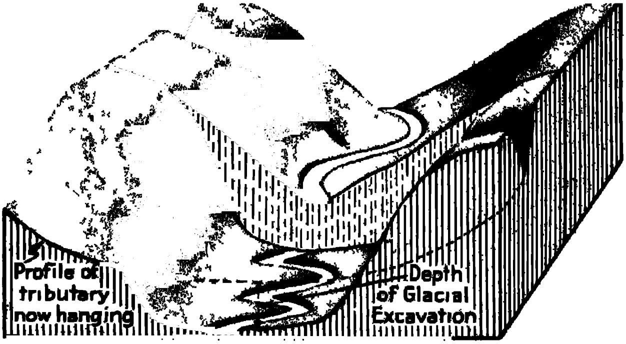


FIG. 119. Block showing alteration of a stream-valley by a valley glacier. Rear half of block shows stream valley before glaciation Front half shows valley deepened and widened by glaciation, with tributaries left hanging above the main valley. (Compare Figs. 118 and 131-134.)

The sculpture of fiords, like the sculpture of other glaciated valleys, is not restricted to valley glaciers. Some fiords have been made by ice sheets as they flowed over mountains or plateaus cut by deep valleys.

GLACIAL TRANSPORT

Movement of rock waste by glaciers is more continuous than movement by streams, because the glacier's load does not so readily sink to the bottom, requiring to be lifted up again. The rock debris carried by ice sheets consists chiefly of material picked up from the ground and is concentrated along the under surfaces of the glaciers. That carried by valley glaciers is augmented by sliderock avalanched from valley walls and is therefore present also at the sides and on the upper surfaces of the glaciers. In both valley glaciers and ice sheets, the great bulk of the ice is commonly nearly free of rock debris. Exceptions are the outer margins of some glaciers, where waste becomes concentrated on their upper surfaces as they shrink by thinning (Fig. 126). A glacier can transport with ease large boulders (Fig. 136) that could not be moved by a stream of comparable discharge. A glacially transported granite boulder near Conway, New Hampshire, measures 90 by 40 by 38 feet; another, of limestone, in eastern Ohio has a surface area of more than one acre.

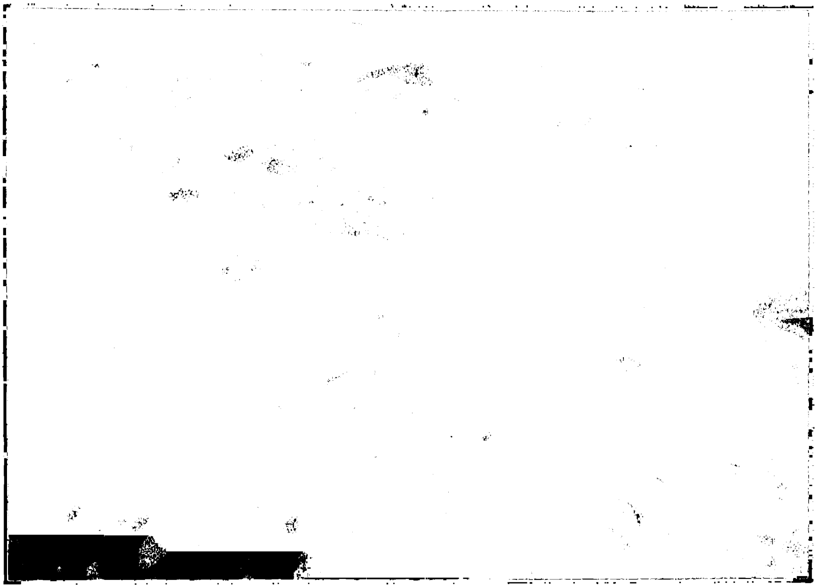
GLACIAL DEPOSITS

Till and Stratified Drift

All the material in transport by glacier ice, and all the material predominantly of glacial origin deposited directly by glaciers or indirectly in glacial streams, glacial lakes, and the sea, together constitute *glacial drift*. The name *drift* dates from a time more than a hundred years ago, when it was conjectured that such deposits had been "drifted" to their resting places by water. In texture and arrangement, drift varies through many gradations between two extremes. One extreme consists of drift deposited directly by the glacier without having been flushed by meltwater. The other extreme consists of drift so thoroughly worked over by meltwater that it has become well stratified and is actually a deposit made by water.

Till. Drift deposited directly by the ice consists of pieces of rock of many different sizes, usually ranging from large boulders down to silt and clay. The pieces are not sorted according to size and weight; they lie just as they came out of the glacier ice (Fig. 120). This kind of drift is known as *till*, a name given it by Scottish peasants long before its true origin was understood. Till comes out of the ice in a variety of ways. Some of it is merely dumped off the ice terminus. Some of it is heaped up, snowplow-fashion, as the ice terminus creeps forward. Some of it is slowly let down to the ground from positions within or upon the glacier, as the ice beneath it slowly wastes away and disappears (Fig. 126). Probably the great bulk of it is plastered on to the ground from the sole of the creeping glacier. Some of the stones in the till—particularly till of the type last mentioned—have glaciated shapes approaching the one shown in Fig. 142, *a*, but the proportion of such stones is commonly very small.

Although ideal till consists of a deposit from which not even the finest particles were washed out during its accumulation, yet in many places the till is more or less washed. This is understandable in view of the fact that, at one time or another, meltwater is present in the terminal zones of nearly all glaciers and is able, by flushing out the finer particles, to modify the initial ice-laid character of the till.



Sanborn Partridge.

FIG 120. Exposure of till, showing its lack of both size sorting and stratification, and the glaciated shapes of some of the stones it contains. (Compare Fig. 121.) Fishell's Brook, St. Georges Bay, Newfoundland

Stratified Drift. Much of the drift is partly or wholly stratified, indicating that meltwater has partly or thoroughly reworked the rock debris carried by the ice (Fig. 121). Stratified drift is complementary to till; the more till, the less stratified drift, depending on the amount and effectiveness of meltwater in rehandling the drift. Stratified drift occurs in immediate contact with the glacier and also is distributed by streams of meltwater far beyond the terminus of the glacier itself. There it is gradually adulterated by stream deposits of local, nonglacial origin, until its identity is lost. The Missouri River carries fine drift from glaciers in the Rocky Mountains of Montana, but, when deposited along the lower Mississippi, this material would be difficult indeed to separate from the other sediments with which it is mixed.

Stratified drift contains few glaciated stones. Boulders, cobbles, and pebbles are successively deposited as the gradient and competence of the meltwater stream diminish. These large pieces are abraded by water-driven "fines"—sand, silt, and smaller particles—and their striations and facets are worn off. The fines, however, are chiefly the product of mechanical grinding by the glacier and are therefore fresh and undecomposed, unlike the fine products of chemical weathering that are delivered by the mass-wasting processes to most nonglacial streams.

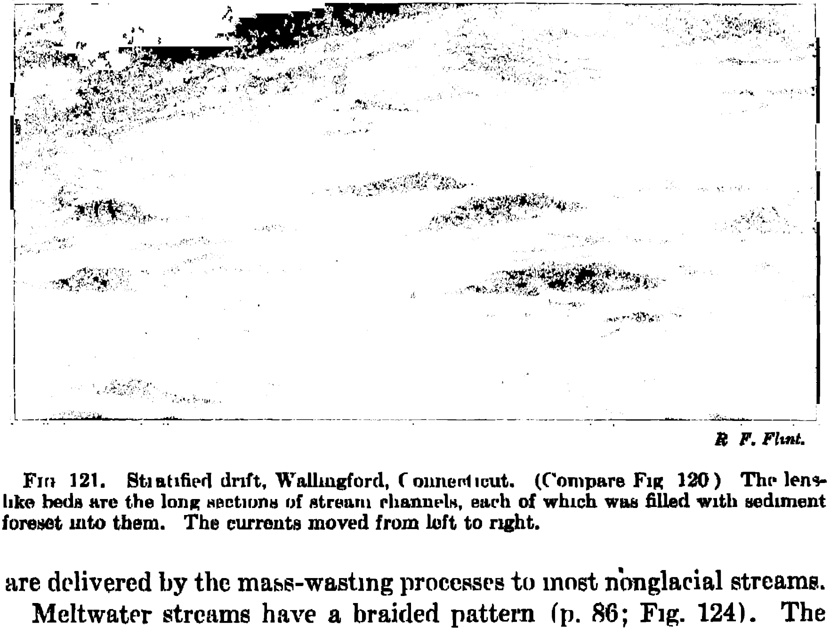

R. F. Flint.

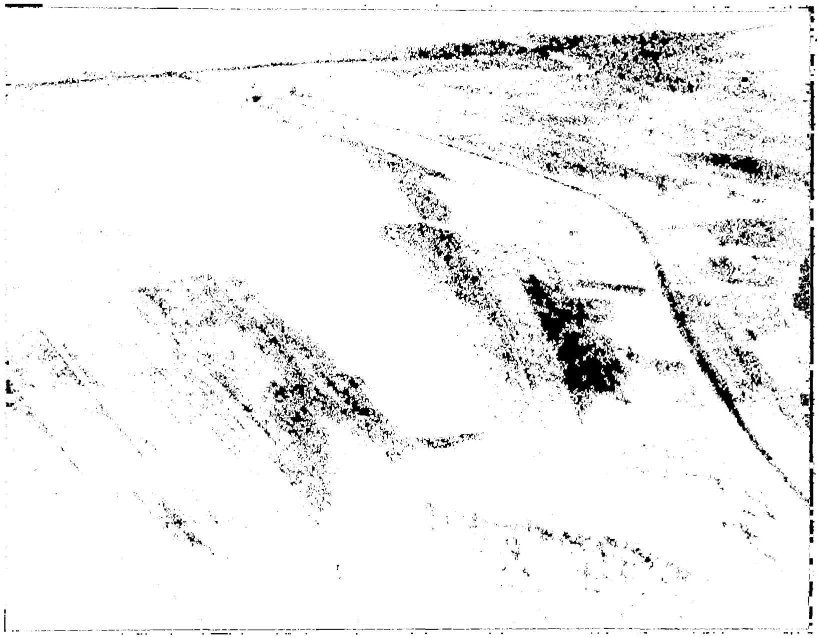
FIG 121. Stratified drift, Wallingford, Connecticut. (Compare Fig 120) The lens-like beds are the long sections of stream channels, each of which was filled with sediment foreset into them. The currents moved from left to right.

Meltwater streams have a braided pattern (p. 86; Fig. 124). The braiding habit results in part from great fluctuations in discharge between day and night, and in part from the fact that the streams are already heavily loaded as they emerge from the glacier

Types of Drift Based on Surface Form

Terms such as *drift* and *till* refer only to deposits, without regard to their surface form. Some glacial deposits, however, including both till and stratified drift, are fashioned into distinctive forms recognizable as landscape features. Some of the best known of these are ground moraine, end moraine, outwash plains, drumlins, eskers, and kame terraces.

Ground Moraine. Drift that is widely distributed, ordinarily relatively thin, consisting chiefly of till, and having a gently irregular initial surface form is referred to as *ground moraine*. The gentle undulations of its surface, in some places including closed depressions, seem to be in part the result of uneven deposition by ice unevenly charged with drift. Ground moraine covers the greater part of the glaciated region from eastern Ohio west to the Rocky Mountains in Montana and is conspicuous likewise in Europe from northern Germany eastward through European Russia. In these regions the drift is thick enough to have obliterated most of the hills and valleys on which it rests (Fig. 130).



Charles C. Bradley.

FIG. 122. Drumlin near Madison, Wisconsin, seen from the air.

Drumlins. The ground moraine in some districts in western New York, central Wisconsin, southern New England, Nova Scotia, and the British Isles is dotted with scores of smooth hills shaped like the inverted bowls of teaspoons. These are *drumlins* (Figs. 122, 123). Most of them are made of clayey till, and many of them have cores of bedrock. Ranging up to more than a mile in length and up to 200 feet in height, their long axes parallel the direction of flow of the former glacier ice. In the ideal drumlin the blunt head opposes the ice flow and the narrower tail points with it. Drumlins are perfectly streamlined forms, offering minimum resistance to ice flow. They occur chiefly where clay is an abundant constituent of the drift and probably owe their origin to a combination of erosion and plastering deposition, the latter process gradually shaping their tails, the cohesive property of the clay playing an essential part Most drumlins are formed not far back from the terminus of the glacier, where the ice is comparatively thin and where drift is abundant. The chief geologic value of drumlins is that they record vigorous ice flow and indicate the direction of movement of former glaciers.

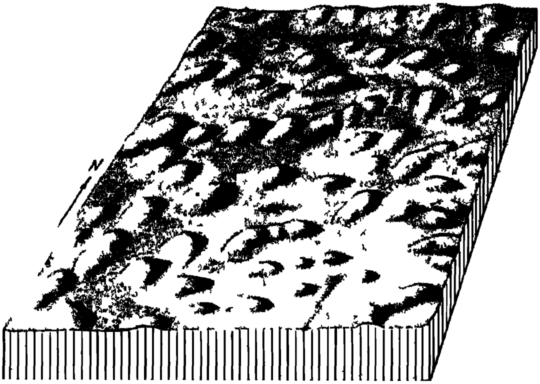


FIG. 123 Relief model of a group of drumlins near Fond du Lac, Wisconsin The area shown is 4 5 by 4 miles. Vertical scale exaggerated

End (Terminal) Moraines. A generally ridge-like accumulation of drift, resulting chiefly from deposition by a glacier along its terminal margin, is an *end* (or *terminal*) *moraine*. Such an accumulation may be made by three processes (p 176) or by combinations of them: (1) the snowplow process in front of an expanding glacier, (2) the plastering process beneath the terminal margin of an active glacier, (3) the dump process at the margin of a glacier whose terminus is stationary owing to temporary equilibrium between flow and wastage

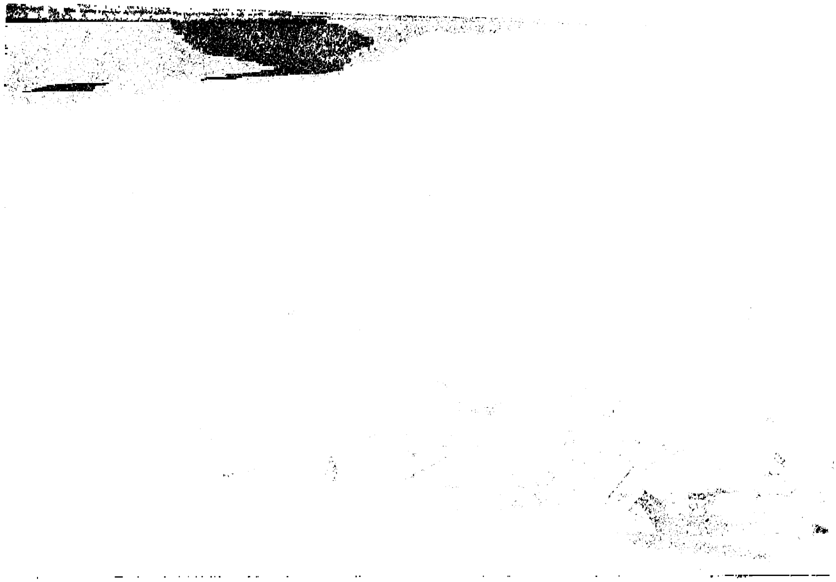
The end moraines of valley glaciers (Figs. 124, 125) range from a few feet to scores and, exceptionally, hundreds of feet in height Ordinarily their breadths are slight. The length of a fully formed and undissected end moraine is controlled by the width of the containing valley. In plan the end moraine is usually convex down-valley, re-

BRADFORD WASHBURN.

Fig. 124. Features of a shrinking valley glacier. End moraines (dark-colored arcs) marking successive positions of the margin of a shrinking valley glacier near Mount Iliamna, Alaska. Braided streams of meltwater have breached the end moraines and are building a great outwash plain of stratified drift. One meltwater stream flows along the lateral margin of the glacier. On the glacier surface are crevasses indicating fracture, and sinuous bands of drift (including lateral moraine at left) indicating flow. The ice has steepened the sides of its valley. In the left background ice-abraded rock surfaces, a rock-basin lake, a hanging tributary valley, and a cirque are visible. (Contrast the stagnant terminal zone of a different glacier, Fig. 126.)

cording more rapid movement of the center than of the sides of the ice.

The end moraines of most valley glaciers contain but a small proportion of the waste removed from cirque and upper valley, much of it having been strewn along the valley floor as ground moraine and much having been carried off down the valley by meltwater. In many steep valleys, indeed, end moraines do not form at all, because the drift is sluiced away by water as rapidly as it forms. The fact that an end moraine of a vanished valley glacier is incomplete, however,



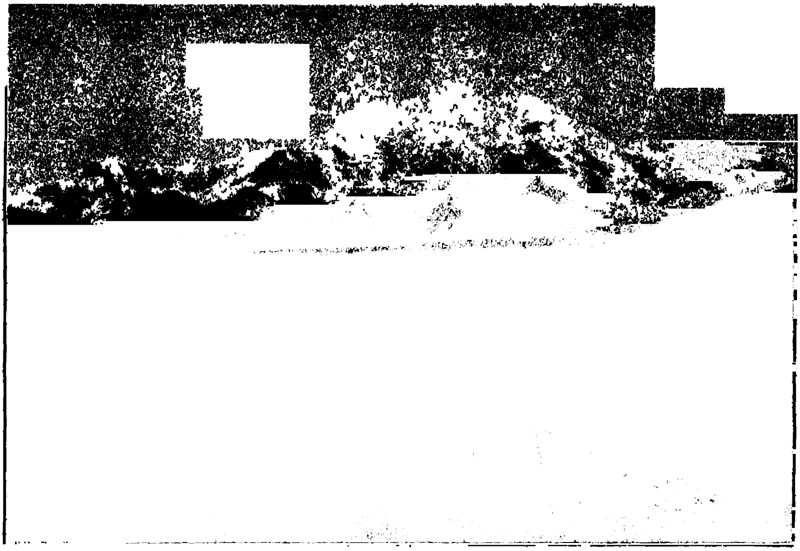
WHITMAN CROSS, U. S. GEOLOGICAL SURVEY.

Fig. 125. End moraines (low ridges in center) and outwash (smooth plain on right) in the Animas Valley near Animas City, Colorado. The ice margin shrank back from right to left, and both moraines and outwash have since been trenched by the river.

does not indicate that it was not once fully formed, as a vigorous stream readily breaches it and eventually destroys it.

End moraines rarely form along the entire margin of an ice sheet because the important factors of topography, atmospheric moisture, and movement of the ice vary greatly from one sector to another. Some of the ridges left at the margin of the ice sheet that recently covered part of North America are scores of miles long, 100 to 200 feet high, and 1 to 5 miles wide. Their slopes are gentle, and except for their broadly ridged character they have almost the same undulatory surfaces that characterize the ground moraine back of them. Their continuity is broken by gaps through which meltwater formerly flowed. End moraines of this kind cross the broad region south of Lake Erie and Lake Michigan in great festoon-like belts trending generally east-west. Each of these belts marks an interruption, probably climatic, of the wastage that caused the ice sheet to disappear. The great Baltic End Moraine in northern Germany is for many miles not a ridge but a trench, because, being rich in boulders, it has been thoroughly excavated as a source of paving blocks and building stone for Berlin and other cities in a region where rock is very scarce.

Superglacial Moraines. Many existing valley glaciers carry long ridges of drift on their lateral margins. Built in part by avalanching from valley sides, these are termed *lateral moraines* (Fig. 124). The merging of two lateral moraines beyond the point at which two valley glaciers coalesce to form a larger glacier produces a *medial moraine*



Alaskan Aerial Survey, U. S. Navy.

FIG. 126. Blanket of superficial drift covering the surface of the stagnant and wasting terminal zone of the Fairweather Glacier, in the Juneau region, Alaska. Most of the drift was contained in the ice and is accumulated on the surface as the ice melts. Kettles and drift knolls are forming on the surface. The underlying ice, streaked with drift, is visible in the cliff in the foreground. The active part of the glacier, upstream from the terminal zone, is seen in the distance as nearly clean ice, flowing toward the observer. The mountain is Mt. Fairweather (15,300 feet).

(Fig. 110). Lateral moraines often survive the glaciers by which they were built, but medial moraines rarely do so, because their contents are spread out during the shrinkage and disappearance of the ice and are lost among the debris accumulating on the surface as wastage thins the glacier (Figs. 126, 127).

Forms Built of Stratified Drift. *Outwash.* A great accumulation of stream-deposited drift in the region beyond the ice terminus is *outwash* (Fig. 124). Choked with drift, the meltwater streams aggrade their beds and shift their braided channels, building up deposits of sediment that may reach thicknesses of hundreds of feet, their chan-

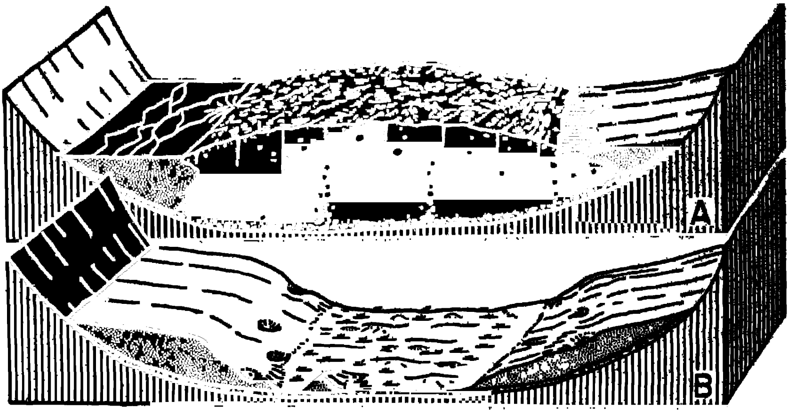
BRADFORD WASHBURN.

Fig. 127. Kettles forming in the drift that covers the nearly stagnant terminal zone of Chitina Glacier, St. Elias Mountains, Alaska. The area of glacier in view exceeds 12 square miles. View down the glacier, past the terminus to the head of the outwash beyond. The small cliffs that inclose the ponds expose glacier ice. (Compare Fig. 126.)

neled upper surfaces sloping away from the ice with steep stream profiles. The traveler along an active outwash plain has a difficult time of it. Because of the braided channel pattern he is nearly always on an island, from which he has no choice but to ford a swift cold stream. Once across the stream he is only on another island.

Some outwash plains, traced headward, lead through gaps in a related end moraine and merge with the ground moraine on the other side. Other outwash masses end headward in steep slopes facing the positions vacated by the former glaciers. The heads of such outwash masses were built against and upon the terminal margins of glaciers which were feebly flowing or entirely stagnant. When the ice wasted away, the stream deposits slumped down, locally reversing their surface slopes. Slopes representing the former contact of drift with ice are *ice-contact faces* (Fig. 128).

Kettles. The ice-contact heads of outwash plains as well as other masses of drift accumulated adjacent to a glacier are commonly pitted with undrained depressions ranging from a few yards up to a mile or more in diameter. The smaller depressions are usually nearly circular; the large ones are more likely to be elongate. Many contain lakes or swamps. Such features are termed *kettles* (Figs. 127, 129), and their origin is well understood because they are forming today near the margins of some glaciers. Masses of glacier ice of various sizes, either surrounded or completely buried by drift, slowly waste away and cause slumping and caving in of the drift.



From Flint, Glacial Geology and the Pleistocene Epoch.

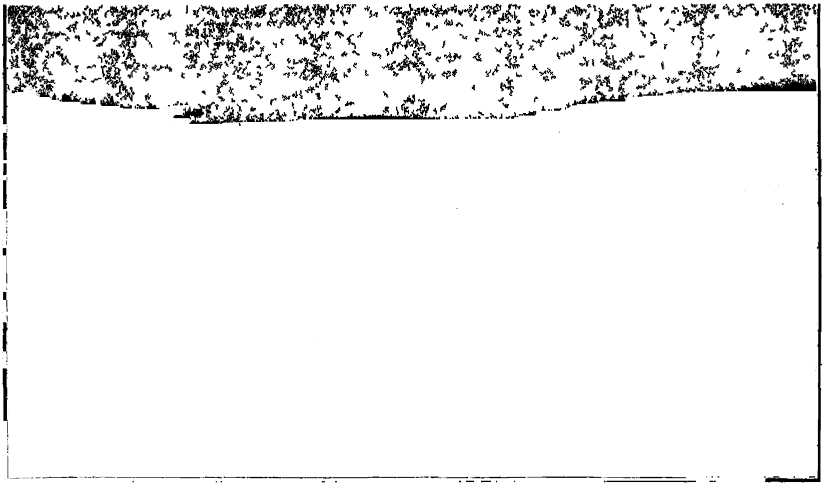
FIG. 128. Making of a kame terrace.

A. Segment of a valley occupied by a shrinking tongue of glacier ice Braided streams are flowing along the margins of the glacier and are building up thick deposits of sediment. A small lake has formed against one side of the glacier.

B When the ice disappeared the marginal deposits of sediment were left as kame terraces (Contrast the terraces shown in Fig. 327, p. 494)

Pitted Outwash Plains. Considerable areas in Minnesota, Wisconsin, Michigan, New England, and northern Europe, and smaller areas in other regions, consist of outwash plains pitted with kettles. The kettles range in diameter from a score of feet up to 10 miles or more and are so closely spaced in some districts that the plain is merely a network of narrow ridges. Such features are convincing evidence of glacier ice that became stagnant. The ice was seamed with a network of crevasses and superglacial stream channels through which ran melt-water, blanketing the ice with stratified drift which ultimately sagged and slumped into its present position.

Kame Terraces. Some glaciated valleys contain terraces made of gravel and sand built up by streams flowing between the side of a glacier and the inclosing valley wall. These are *kame terraces* (Fig. 128); some of them have ice-contact faces. Kame terraces differ both in form and in origin from terraces that have been carved by streams from deposits that formerly filled the valley from side to side. Ordinarily the ice-contact faces are very irregular, with bowl-like indentations, and finger-like projections that have been called *crevasse fillings*. Near these faces the terrace surfaces contain kettles, and in some places isolated knolls of sand and gravel dot the low area beyond the terrace faces.



H T Nation, British Columbia Bureau of Mines.

FIG. 129. Kettles in kame terrace, Chilko River valley, Lillooet District, British Columbia.

Terraces having these features indicate that the ice against and upon which they were built must have been nearly or quite stagnant. Terraces otherwise similar, but lacking these details, were built along the margins of glaciers that probably were actively flowing.

Eskers. In New England, southern Michigan, and Minnesota, and less conspicuously elsewhere in North America, there are winding ridges of stream-built stratified drift, which reach lengths of many miles and heights up to 60 feet or more. The ideal form has steep sides and a narrow top. In North America, such features are called *eskers* (Fig. 130), although the word is used somewhat differently in other

W. S. COOPER.

Fig. 130. Esker resting on ground moraine, Morrison County, Minnesota. View from the air. The former ice sheet flowed toward the observer.

countries. This term is believed to have been derived from an Irish word as old as the Christian era, signifying a ridge or path. Eskers are the most conspicuous features of the landscape in parts of central Ireland, where they were used as footpaths by early inhabitants because they stood above the adjacent boggy lowlands.

Probably these curious forms originated in several ways. Some of them were formed in subglacial tunnels; others were built in crevasses without roofs; still others were built successively at the mouths of subglacial streams, where they emerged at the ice terminus into glacial lakes or the sea and were added to yearly as the terminus migrated backward by calving. Most of them appear to indicate that the inclosing ice had become stagnant or nearly so before the eskers were built.

Records of Active and Stagnant Ice. The glacial forms described above fall into three general groups: (1) ground moraine (locally with

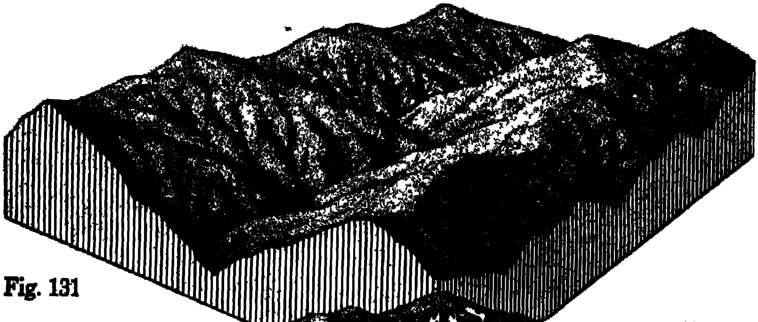


Fig. 131

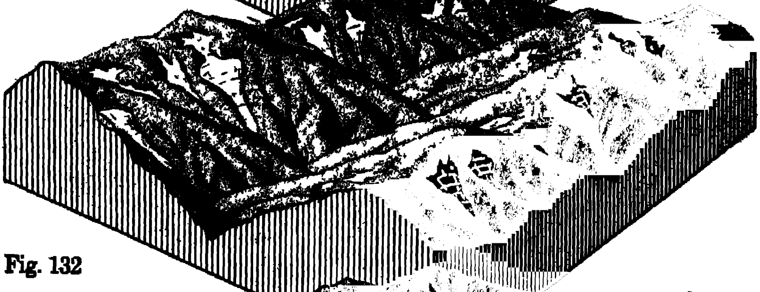


Fig. 132

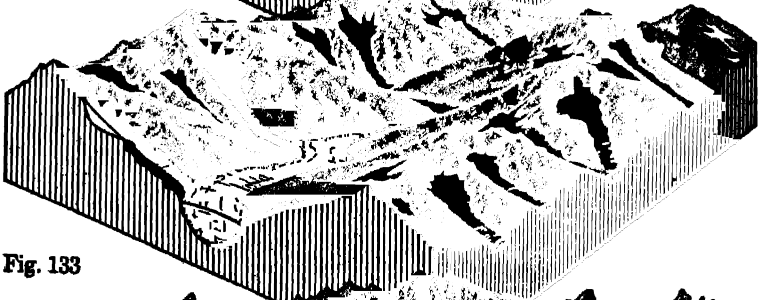


Fig. 133

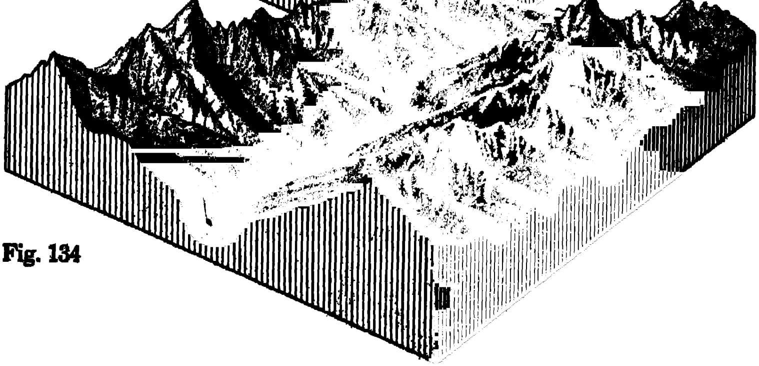


Fig. 134

drumlins) and end moraines, which record actively flowing ice during their building; (2) pitted outwash plains, some kame terraces, large kettles, and eskers, which record stagnant ice during their building; (3) nonpitted outwash plains, which, having been built beyond the ice terminus, do not fall into either group, although they may be associated with either.

As a rule, (1) and (2) do not occur in the same district. From this we may infer that the physical conditions of former glaciers varied from time to time and from place to place. It is likely that the passage of the terminal part of a glacier from an active to a stagnant condition was brought about by two factors. The first was thinning of the ice by decrease in precipitation relative to wastage. The second was increase in frictional resistance to glacier flow because of either rough topography beneath the glacier, or a heavy load of drift in the ice, or both.

THE GLACIAL AGES

Early in the nineteenth century, naturalists in Switzerland began to notice that the boulders and other deposits lying in the ice-free valleys of the Alps were identical in shape and arrangement with the deposits being made by living glaciers in the ice-filled valleys. They realized that the glaciers of the Alps, high and comparatively small then as now, had at some former time spread outward and downward, filling most of the valleys of Switzerland with ice. Carrying this inference still further, Louis Agassiz, a paleontologist, in 1837 announced his opinion that former glaciers had covered not only Switzerland but a large part of Europe. In publishing this bold idea he paved the way for the now universal belief that the climates of the recent geologic past permitted valley glaciers to sculpture many mountain ranges and permitted ice sheets to overflow vast areas in North America, northern Eurasia, and southern South America. At one time these glaciers to-

FIGS. 131–134. Erosion of mountains by valley glaciers.

FIG. 131. A mountainous region being eroded by streams. The main valley is flanked by interlocking spurs.

FIG. 132. Evolution of cirques as the climate grows colder and snowfields become perennial. The snowfields in the cirques generate small valley glaciers.

FIG. 133. Coalescence of the small glaciers to form a large ice tongue in the main valley, enlargement of the cirques, and frost sculpture of the uplands.

FIG. 134. Appearance of the mass after further glaciation, followed by complete disappearance of the glaciers. The valleys have been deepened and widened, the interlocking spurs reduced, the tributaries have been left hanging above the main streams, and rows of empty cirques, some containing lakes, scallop the uplands, which in turn have been converted into knife-edged ridges punctuated by pyramidal peaks.

gether covered more than one-third of the land area of the globe. The glacial invasions occurred repeatedly throughout a total span of more than a million years, and the cold times are referred to as *glacial ages.*

Since the time of these pioneer observations, the work of the vanished glaciers has been closely studied. The general results of the study are summarized in the following sections.

Forms Fashioned by Valley Glaciers. Landscapes fashioned by valley glaciers in mountains like the Alps are distinct from landscapes modeled by ice sheets. Their essential features are shown in Figs. 131–134. Figure 131 represents a mountain region sculptured by streams In Fig. 132 the climate has become colder, snowfields have accumulated in high protected places, and small glaciers have begun to flow out of the snowfields and down the stream valleys. In Fig. 133 the glaciers have expanded to the maximum size permitted by the colder climate. They have widened and deepened the valleys and have left the tributaries hanging above the main valleys. At the same time they have beveled off the ends of lateral spurs, thereby straightening the main valleys. The enlargement of cirques on opposite sides of a ridge has reduced the ridge to a knife edge, which is kept sharp by frost wedging. Two opposing cirques, back to back, slice through the ridge to form a broad gap, and, where several cirques surround the base of a high peak, the peak takes on the shape of a pyramid, such as the Matterhorn in the Alps or the Grand Teton in Wyoming.

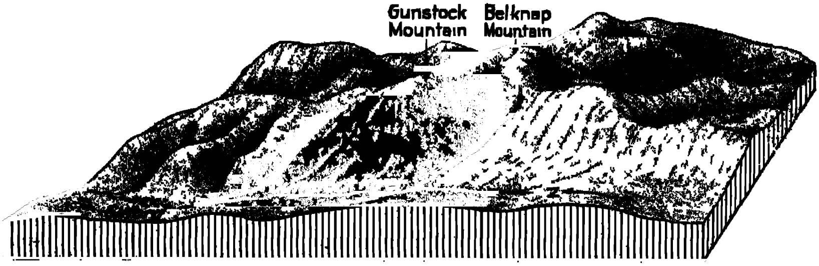
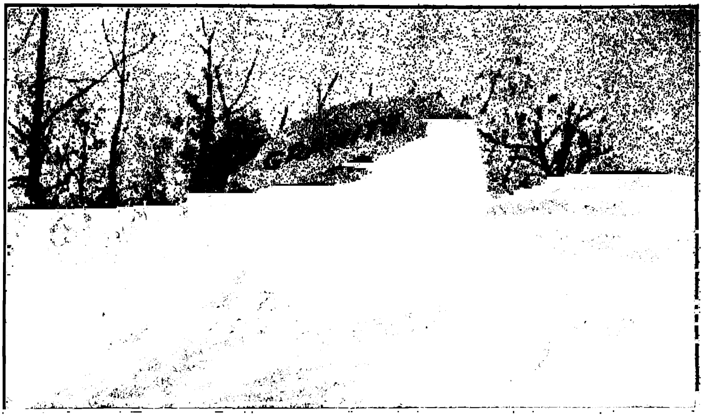


FIG. 135. Belknap Mountains near Lake Winnepesaukee, New Hampshire, showing profiles smoothed by ice sheets. Area of block, 5 x 4 miles. Maximum relief, 1800 feet. Vertical exaggeration, 2 times.

In Fig. 134 the climate has grown warmer, and the glaciers and snowfields have wasted away, leaving straight U-valleys, hanging tributaries, cirques, and jagged sharp-crested ridges that contrast sharply with the smoother mountains of Fig. 131. Most of the very high mountains of the world exhibit these features.

The Former Great Ice Sheets. The New England-Adirondack region, the highlands of Scotland and Ireland, and the interior of British Columbia are mountainous and, except for some of the highest peaks, have been recently more than once glaciated; yet their landscapes differ strikingly from that described in the foregoing account. The crests of the ranges (Fig. 135) are not sharp; there are no knife-edge ridges and no pyramid-like peaks. Cirques are very rare, and the few that are visible are poorly developed. On the contrary, even the ridges are smoothed and rounded, and striated bedrock surfaces are present on them as well as in the valleys. In any one district the striae on all the ridges point in a single general direction. Evidently glaciation here did not involve frost wedging and the excavation of cirques but consisted chiefly of quarrying and abrasion. It is clear that these regions were affected not by mere tongues of ice in their valleys, but by huge ice sheets thousands of feet thick that overtopped valley and range alike, as do the ice sheets of Greenland and the Antarctic Continent today.



C. R. Longwell.

FIG. 136. Angular ice-transported erratic boulder of granite perched on the top of a high ridge of dolerite, Mount Tom, near Northampton, Massachusetts. The nearest outcrops of granite lie many miles away.

The area covered by the former ice sheets is determined by tracing the limits of the glaciated regions. The directions of flow are identified by tracing the paths followed by stray stones and boulders (*glacial erratics*, Fig. 136) derived from known sources in the bedrock. Further evidence is obtained from the directions of grooves and striations, the

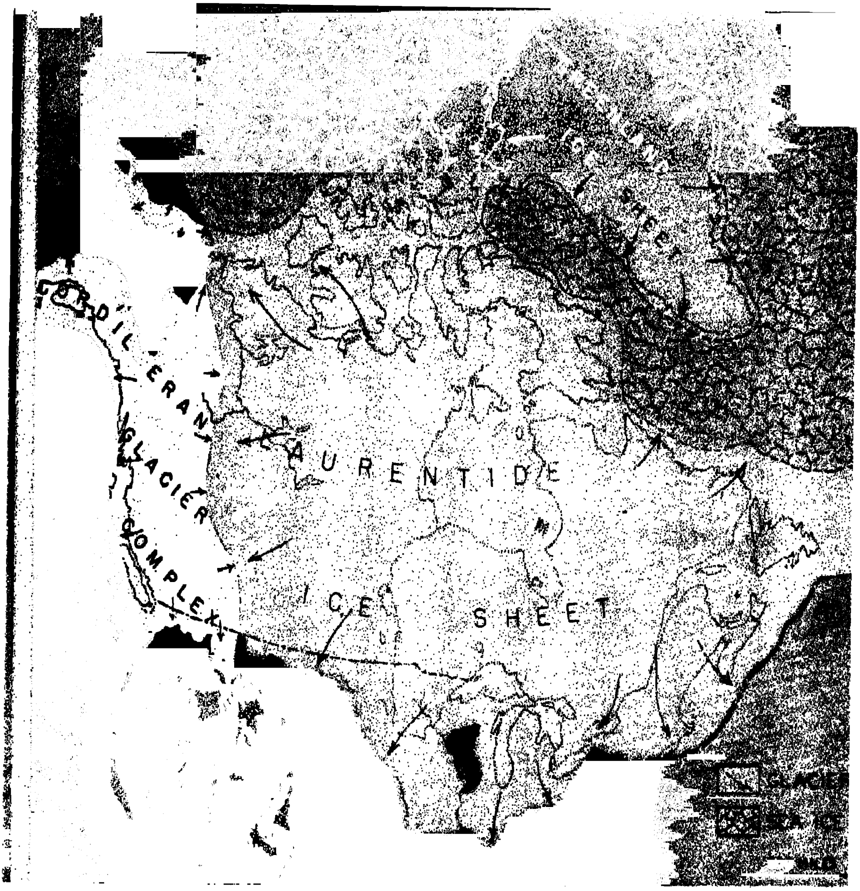


Fig. 137. Extent of glaciation of North America at the maximum of the glacial ages. Somewhat generalized, especially in Arctic Canada, on which information is very scanty. The boundaries between glacier ice, sea ice, and open sea are conjectural, but are based on modern analogies. Arrows show generalized directions of glacier flow. Note relation of Ohio and Missouri rivers to former ice margins.

long axes of drumlins, and the general trends of eskers. The results for North America are shown on the generalized map, Fig. 137.

By matching the drift with the bedrocks from which it was transported, it has been determined that the former glaciers of North America consisted of two distinct masses that originated in two different regions. The largest and most conspicuous mass originated in northern and eastern North America. It grew into a vast ice sheet that spread northward into the Arctic Sea, and southward and westward as far as New Jersey, Pennsylvania, the Ohio River, the Missouri River, and the east base of the Rocky Mountains. When at its maximum, its area probably exceeded 5,000,000 square miles, that of the existing Antarctic Ice Sheet. The lesser mass of glacier ice formed in the mountains of western North America. It included valley glaciers and piedmont glaciers that thickened and coalesced repeatedly, at times forming a virtual ice sheet in the region between the Canadian Rocky Mountains and the Coast Ranges. This confluent mass of glaciers was 1200 miles long and 250 to 400 miles wide. South of it, in western United States, there were more than seventy separate areas of glaciers, some of them large, occupying the higher mountain ranges. Very small glaciers formed even as far south as southeastern New Mexico and southern California near Los Angeles.

In Europe, the principal ice sheet formed in Scandinavia and spread south and east as far as England, the Netherlands, central Germany, Poland, and southern Russia. Many mountain ranges in both Europe and North America supported large valley glaciers, some of which still exist, although greatly diminished in size.

Direct Effects of Former Ice Sheets. The effects of the vanished ice sheets are considerable. Throughout a huge area in Canada north of the Great Lakes, the weathered mantle was stripped away, and the ice bit into the fresh bedrock beneath, so that today it is naked or thinly covered with drift. Many of the thousands of lakes in that vast region lie in shallow basins scooped out of the rock by the ice.

Nearer the terminal zone the drift is more abundant. Although most of the material of the drift has not been moved more than a few miles, some of it has traveled many hundreds of miles. This is proved by the finding of rocks far distant from their places of origin. Slabs of native copper from the great copper lodes of the Keweenaw Peninsula in Michigan have been found as far south as Missouri, and boulders of an unusual conglomerate containing reddish pebbles of jasper, found throughout Ohio, have been traced to outcrops of bedrock on the north shore of Georgian Bay in Canada. In eastern Finland, boulders of ice-transported copper ore were traced backward along the line of former ice flow. This resulted in the discovery in 1910 of Finland's most important copper mine. Several isolated diamonds of good quality have been found in glacial deposits in Wisconsin, Michigan, Ohio, and Indiana. Their position in the deposits indicated clearly that they had been transported and deposited by the ice, but their source is as yet unknown. It is inferred to be somewhere in central Canada.

The surfaces of New England, eastern New York, and the Scottish Highlands are strewn with glacial boulders (Fig. 136). In these regions of resistant rocks the former ice sheets found it difficult to grind into sand and silt the fragments of bedrock they acquired. From Ohio to the Rocky Mountains and throughout northern Germany, however, the much softer bedrock was more easily ground up by the ice, and therefore in these regions large boulders are rare, those found being strays (*erratics*) from regions of harder rocks far to the north.

Influence of Former Ice Sheets on Drainage. Many of the streams flowing away from the glaciated region were choked with glacial sediments and built great deposits of outwash in their valleys. Parts of these deposits still remain; those made by meltwater that came from the latest of the ice sheets and flowed down the Mississippi valley have been traced down to the region of the river's mouth. Many valleys draining toward the expanding glaciers were filled with meltwater as the ice blocked their mouths, and in consequence numerous lakes were formed. Most of these lakes were drained when the glaciers shrank away, but some of the largest of them, including the present Great Lakes, still remain, partly because the flowing ice had scooped out the preglacial valley floors and thus formed rock basins, and partly because of blockades of drift not yet cleared away. As the ice shrank, it uncovered successively lower points in the basin rims and thereby caused the lake levels to fall. Even though the ice has disappeared, the lake water is still spilling over basin rims in several places, the most notable of which is Niagara Falls (pp. 83, 84). The former, higher lake levels are still recorded in old shorelines, beaches, lake deposits, and outlet channels.

The Ohio River throughout the greater part of its length is chiefly a product of glaciation. Prior to the advance of ice sheets into these latitudes, streams in West Virginia and Kentucky drained northwest and north across Ohio and Indiana. The ice sheet flowing from the Labrador Peninsula blocked the valleys and ponded the streams. The rising lakes overflowed westward across the series of divides that lay nearest to the glacier margin, and the overflows were held in these positions long enough so that they were enabled to cut, along the ice margin, a single valley which was followed by the regional drainage even after the ice sheet had disappeared (Fig. 137). This valley is the valley of the Ohio River.

The Missouri River in the Dakotas had a similar history. Prior to the glacial ages the lower Missouri in Missouri and the upper Missouri in Montana were independent streams. An ice sheet flowing southwest from the region west of Hudson Bay blocked the normal easterly drainage of the intervening area and forced it to flow south along the glacier margin, thus integrating the two rivers into the single stream of today (Fig. 137).

The great former lakes Bonneville and Lahontan (p. 150) as well as scores of smaller lakes that formerly dotted the desert basins of western United States were brought into existence by the relatively cool moist climates of the glacial ages.

Effect of Glaciers on the Level of the Sea. Whatever the cause of glaciation, we have no doubt as to the general effect on the sea brought about by the building of glaciers on the lands. Ice sheets are built of atmospheric moisture precipitated chiefly in the form of snow. All this moisture is derived ultimately by evaporation from the sea. It follows that the greater the amount of moisture locked up on land in the form of snow and ice, the less water there is left in the sea. Although the area of the sea outranks that of the land by a ratio of about 3 to 1, nevertheless it is estimated that the complete wastage of all the glacier ice existing today would return enough water to the sea to raise its level about 100 feet. This would drown vast areas of land, much of it densely populated, and would submerge large parts of such cities as New York, Boston, London, and Hamburg. But the glaciers formed during recent geologic history covered nearly three times as great an area as do the glaciers of today. In consequence the level of the sea at those former times must have been considerably lower than it is now—perhaps by as much as 300 feet.

Accurate measurements made since 1850 strongly suggest that the sealevel is now rising at a rate of about 2.5 inches per century. This rise is a direct consequence of the rapid wastage of the world's glaciers (p. 164) that has been in progress during the same period.

It is evident therefore that glaciation has not only an important direct effect in sculpturing the land, but also an important indirect effect in shifting the line of attack of the sea upon the land (Chap. 11) and thus in influencing erosion both by the sea and by streams.

Effect of the Weight of Glaciers on the Earth's Crust. As the ice sheets formed and thickened, their weight depressed the Earth's crust beneath them, and as they disappeared the crust recoiled by a slow warping movement that is still in progress at the present time. The features on which our knowledge of these movements is based, including marine deposits now above sealevel and shorelines (p. 228) formed upon the downwarped crust and later deformed out of their initial horizontality, are discussed more fully on page 356.

Repeated Glacial Ages. The glacial deposits of both North America and Europe consist of distinct layers of drift separated from each other by thin and very discontinuous deposits made under ice-free conditions. This evidence indicates that there were no fewer than four successive incursions of ice separated by *interglacial* times when climates were warmer. During at least one of these times the ice may have almost wholly disappeared, under a climate somewhat warmer than the present one, as indicated by fossils contained in the deposits.

Glaciations in More Remote Geologic History. The slightly weathered character of the youngest drift indicates that the latest extensive glaciers in North America and Europe wasted away so recently that the date of their disappearance may be reckoned in mere thousands of years. Likewise the condition of the older drifts implies that the four successive ages, as a group, date probably from within the last million years or more. These deposits, however, are comparatively very recent. Occasionally there are discovered layers of rock, scores and even hundreds of millions of years old, that prove to be solidified till containing glaciated stones. Because these very ancient glacial deposits have been found in many parts of the world and in rocks of wholly different ages, we conclude that glaciation has occurred at several times during the remote past, separated by very long intervals during which there may have been no ice at all on the Earth, even near the poles. Therefore the study of glacial features is important in that it enables us to recognize them in deposits of various dates, and thus to add to our knowledge of the changing climates of the past.

Cause of Glacial Climates. The mean temperature that characterized the climates of the glacial ages in middle latitudes was perhaps 8°C. lower than the mean temperatures now prevailing in these latitudes, and the interglacial temperatures may have been 2 or 3 degrees warmer than now. The cause of these fluctuations of temperature has not yet been established. Several explanations have been put forward, but most of them do not explain all the facts. The explanation that seems best to fit the facts now known involves two unrelated factors—topography and solar heat.

Topography. Glaciers are closely related to lofty highlands. The last million years, embracing the repeated glacial ages, have constituted a time of far higher lands than have characterized the Earth throughout most of its history. This span of time has witnessed the culmination, not only of broad elevation of the continents, but of the formation and upheaval of many new mountain ranges in various parts of the world. Furthermore the geologic evidence shows that the earlier glacial times, dating back to remote periods in the history of the Earth, were likewise times of high lands and conspicuous mountains. In between these times, altitudes were comparatively low and seas were widespread. Highlands are favorable to glaciation for various reasons. Broad uplifts of the continents increase climatic contrasts, reducing temperatures over the lands. The rise of mountain ranges lowers temperatures at their crests and erects barriers to movements of the atmosphere. The latter effect reduces the atmospheric transmission of heat from low latitudes to high, localizes precipitation, and creates clouds which reflect solar radiation and thus lower the temperature at the Earth's surface. Thus it appears that without highlands, extensive glaciers could scarcely take form.

Solar Heat. To explain the glacial and interglacial climates some factor other than the rise of highlands is required, for there is ample evidence that the lands did not rise and sink; they stood generally firm while the climates changed. This looked-for factor may consist of fluctuation in the amount of heat emitted by the Sun. Direct measurement has shown that this amount currently fluctuates, at irregular intervals, through as much as 3 per cent of its average value. In order to create extensive glaciers a larger fluctuation would be required, but whether such larger fluctuations have actually taken place is entirely a matter of speculation. Astronomic evidence neither confirms nor denies them. The best we can say is that, given the highlands, larger fluctuations in solar radiation apparently could have brought about the glacial and interglacial climates that are clearly recorded by the geologic evidence.

Once a large glacier is formed, it is to some degree self-extending, chiefly through the creation of winds that flow outward beyond it, reducing temperatures and paving the way for glacier expansion, but also through the reflection of solar heat by the white surface of the ice—a factor by no means negligible. In consequence, comparatively small reductions in the present mean annual temperatures of northern Europe and North America would be sufficient to create glaciers that in time might expand to large size.

Frozen Ground. In temperate latitudes the moisture in the uppermost few inches or few feet of the mantle freezes during the winter, cementing the ground into a hard mass which thaws out and loosens again in the spring. But in vast areas throughout high latitudes the ground is deeply frozen perennially. In some districts the depth of freezing amounts to many hundreds of feet. The summer warmth thaws the uppermost few feet of ground, but with the coming of autumn the thawed part refreezes.

Perennially frozen ground occurs in regions where the mean annual temperature is several degrees below the freezing point. Its presence introduces serious complications into the problems of constructing buildings, highways, and railroads, as residents of Alaska, Siberia, and other high-latitude lands have learned by sometimes disastrous experience. In addition to its practical, engineering interest, frozen ground is of interest to the geologist in another way. Repeated superficial thawing and refreezing of deeply frozen ground rearranges the constituents of the mantle in a recognizable way. The effect of this rearrangement is seen today in localities in temperate latitudes, proving that the mantle in those localities was formerly perennially frozen—unquestionably during the glacial ages.

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CHAPTER 10

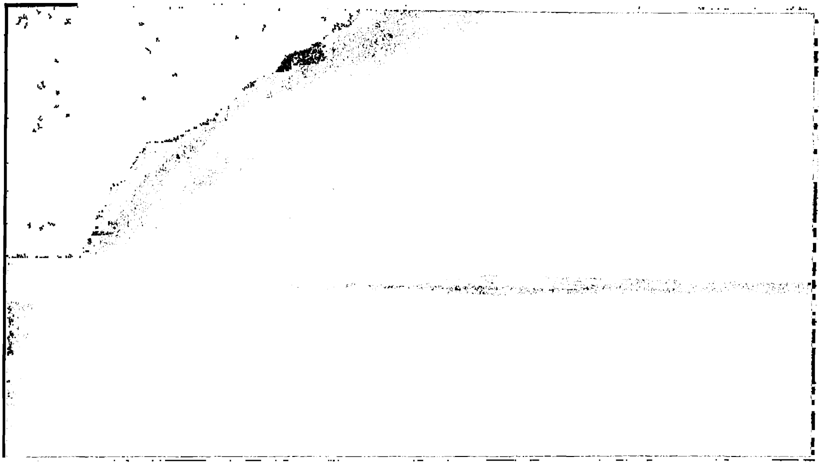
EROSION AND DEPOSITION BY WIND

Moving currents of air have both indirect and direct geologic effects. They are of prime importance in controlling weather conditions, including rainfall; and these conditions in turn govern not only weathering of the rocks but also erosion by streams. Winds also create waves and currents on the sea; thus they bring about marine erosion of coasts and also establish a large part of the mechanism by which sediments derived from the lands are distributed on the sea floor (Chap. 11). If the atmosphere in motion performed only these indirect functions it would be one of the most significant geological agents; but its importance is increased greatly by its power directly to erode and distribute rock material.

WIND EROSION

Erosion by wind is of two distinct kinds. Loose particles of the mantle, such as grains of sand and silt, are picked up by moving air and carried from one place to another. This process is *deflation* (from Latin *de* + *flare,* to blow away). In their motion the wind-driven particles strike against each other, against pebbles and boulders on the ground, and against exposed bedrock; as a result additional particles are worn and chipped from the bedrock and from the individual loose pieces. This process, analogous to abrasion by running water (p. 77), is *wind abrasion.*

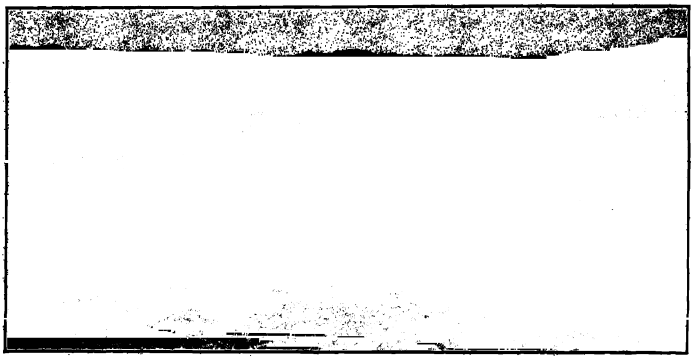
Deflation. In regions that have large or moderate rainfall the results of deflation are not conspicuous. Grass and other vegetation protect a large part of the mantle, and where no vegetation exists the soil particles and sand grains are held together by moisture much of the time. In dry weather, however, clouds of dust are raised from streets, roads, and plowed fields, and the abundance of fine material transported by a hard dry wind is suggested by the quantity swept into houses. Along seashores the beach sand dries to some extent at low tide, and the wind blows it bit by bit beyond the reach of high tide. Storm waves and unusual tides also carry quantities of the sand to a high level, where it becomes thoroughly dry and is driven farther inland by onshore winds. The shores of large lakes also are favorable localities for wind work; parts of Lake Michigan, for example, are bordered by large areas of shifting sand. But the essential combination of conditions for effective deflation—abundant fine-grained, dry mantle unprotected by vegetation—exists only rarely and locally in lands with humid climates. Not only is the ground frequently dampened by rain, but also the water table is near the surface, and much of the mantle is kept moist by capillary action (Chap. 7). In semiarid and arid regions vegetation is scanty, and the upper part of the mantle is dry most of the time. As fine particles are detached by weathering they are caught up and moved by wind; this action explains in part the abundance of rock outcrops in dry regions of rough topography. Silt and sand carried into lowlands by streams are shifted in large quantities by the wind (Fig. 102, p. 149). Even when the weather is comparatively quiet, the air, heated by contact with the hot ground, rises in whirls and lifts the dust in tall columns which move slowly across the plains. On a hot summer afternoon dozens of these columns can be seen at the same time in different parts of a wide arid basin. During storms the air is filled with dust to a great height, and a sheet of sand is driven along the ground (Fig. 138). In the deserts of central Asia and Africa sand storms are a great danger to travelers.



G. N. Morhig.

FIG. 138. A sandstorm sweeping over Khartoum North, Anglo-Egyptian Sudan. Blue Nile River in foreground.

Large quantities of fine debris driven out of the Sahara by exceptional winds fall in the countries of southern Europe, and also in a wide area of the sea as evidenced by dustfalls on the decks of ships in the Mediterranean and hundreds of miles off the west coast of Africa. Great areas in the Libyan Desert of northern Africa are littered with boulders and pebbles that once were scattered through a considerable thickness of finer-grained loose material. When the strong winds blew away the surrounding silt and sand the coarse fragments remained in place, except that they settled down slowly as they were undermined, and so they have accumulated until they mantle the entire surface.



Eliot Blackwelder.

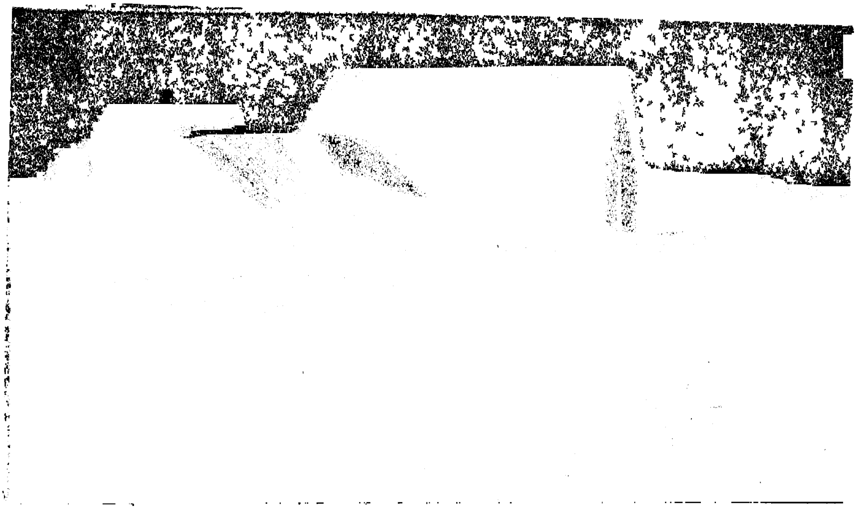
FIG. 139. Danby Playa, Mohave Desert, California. The low hills are capped with a layer of gypsum, which gives the underlying silt some protection from erosion. Presumably the caps were once continuous, and all the material that once filled the spaces between the buttes has been blown away by wind. (Since the playa is in the lowest part of an interior basin, nothing can be carried out by running water.)

The effects of deflation are conspicuous also in some parts of the Mohave Desert and other arid regions in southwestern United States. Sand storms are common in the basins (p. 116), and every strong wind sweeps clouds of dust from the bare surfaces of playas (Fig. 102, p. 149). Remnants of layers that once were continuous over some of the playas now form isolated knolls, the heights of which give a minimum measure of erosion accomplished by the wind within comparatively recent times (Fig. 139). Numerous shallow, undrained depressions in arid and semiarid lands clearly have resulted from localized deflation. Many gentle slopes above the levels of playas are veneered with "desert pavements" consisting of pebbles fitted so closely together and with their top surfaces so even that the general effect suggests a mosaic.

Such surfaces result from slow removal of the fine material, partly by deflation and partly by slope wash, until the pebbles are concentrated to form a continuous layer and thus the fine material beneath is protected from further erosion.

No doubt much of the silt blown from a desert lowland is dropped in other parts of the same basin or in neighboring basins and later is returned by storm waters to the playas. However, the wind carries much of the finest material across the mountains and drops it outside the region of interior drainage. In this way the Gobi Desert and other arid basins of central Asia have lost vast quantities of fine silt, part of which has accumulated as a thick cover on the hills and plains of northern China (p. 220). It is not possible to estimate the rate at which the average surface of an arid region is lowered by deflation, but the process appears to be extremely slow in comparison with stream erosion in regions of plentiful rainfall. Even in arid lands streams may accomplish considerably more erosion than the wind. It should be kept in mind, however, that in areas of interior drainage, such as the Great Basin, running water can do no more than level the surface in some degree by eroding material from higher altitudes and depositing it in the basins (p. 115). The only actual lowering of the general surface through erosion is accomplished by the wind, which is the one agent that can carry material outside the boundaries of the region.

Extreme aridity is not essential to effective deflation. On the Great Plains east of the Rocky Mountains and in similar semiarid districts, large quantities of the silt and sand deposited by streams during floods are blown from the dry floodplains and channels in times of drought when the streams shrink to small trickles or disappear. Not all parts of the uplands are protected by grass, and the wind takes its toll where dry soil is exposed. Many people in eastern United States were not aware of the existence of the "Dust Bowl" until several years of exceptional drought in the Great Plains region, starting about 1930, resulted in dust storms of unusual severity. Such storms have always been common in that region; but plowing up of the sod in large areas during and after World War I, in response to the large foreign demand for wheat, exposed great quantities of loose soil to wind action. Overgrazing of range lands in the semiarid belt has had a similar effect. During the most violent storms, known locally as "black blizzards," the clouds of dust cause almost total darkness at midday. Millions of tons of soil are moved in a single storm; some of it accumulates in local drifts that almost bury farm buildings (Fig. 140), but vast quantities travel hundreds of miles in the upper air, to sift down over wide areas in the



U S SOIL CONSERVATION SERVICE

Fig 140 Wind-blown sediment, chiefly silt, piled up against farm buildings. The roof of the nearest building has collapsed under the weight. Tripp County, South Dakota

Mississippi Valley. Dense haze caused by fine dust from some of the worst storms has been noticeable as far east as New England.

Wind Abrasion. The abrasive effect of wind-blown sand is illustrated by the artificial sandblast, operated with compressed air, which is used to clean the begrimed surfaces of stone and brick buildings. Along coasts the sand driven by winds abrades so effectively that glass in the window panes of houses has been known to lose its transparency during a single storm. In some arid regions wooden telegraph poles can not be used because they are cut down by wind-blown sand, and in exceptional localities the steel rails of railways have been worn thin by the natural sandblast. The effectiveness of wind-blown sand as a cutting or abrading agent is explained by the hardness and strength of sand grains, most of which are made of quartz. Any grains composed of soft minerals like calcite, or of minerals such as feldspar that are weakened by cleavage planes, are ground or broken to bits by continued wear, and the quartz grains, with some garnet and other less common hard minerals, become concentrated. Thus, paradoxically, with continued use the tools of wind abrasion may become more effective.

The most essential conditions for wind abrasion on a large scale are extreme aridity with consequent lack of vegetation, persistent strong winds, an abundance of hard sand grains, and bedrock that is fairly soft or weakly cemented. These conditions are combined ideally in the Libyan Desert, in northern Africa west of the Nile Valley, one of the most rainless districts known. For eleven consecutive years the weather station at Dakhla reported no rainfall. Except for a few small scattered oases around springs and wells the country is almost without vegetation. Strong winds blow steadily from the northwest, and since there are no abrupt mountain ridges the force of the wind is not seriously checked. In the northern part of the region there are large outcrops of weakly cemented sandstone which disintegrates easily; the loose sand thus formed is carried to the south, where the bedrock is chiefly limestone and weak sandstone. With such a combination of favorable factors, wind abrasion is an important process locally in wearing away the surface of the land. Its effects are most evident on the limestone, which is soft but firm and compact, and therefore becomes polished and grooved under the persistent action of the sandblast (Fig. 141). Hard objects in the rock, such as fossils or flint nodules, are brought into strong relief as the surrounding limestone is etched away, and outcrops are carved into rugged fantastic forms. Since abrasion is most effective near the ground, where most of the coarse

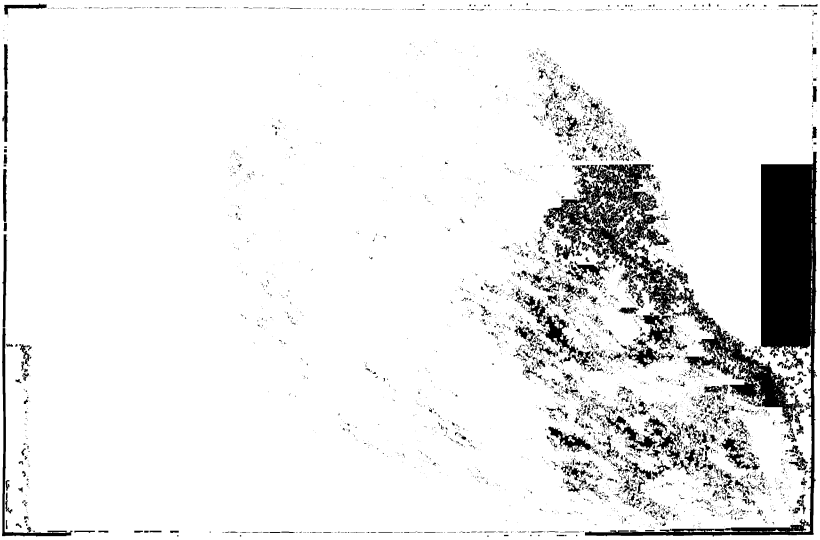

C. R. Longwell.

FIG. 141. A block of dolomite abraded by wind-blown sand. Arrow shows general direction of wind. The face *F* of the block was protected from abrasion by a covering of silt and gravel in which the block was partly buried. Note the grooving of the abraded face parallel to wind direction (width of block about 6 inches).

sand moves, cliffs tend to be undercut, and slender columns are worn at their bases until they topple over. Areas that have unusually weak bedrock and others that are most favorably situated for wind attack are abraded, and the ground-up material is blown away until large undrained depressions are formed. None of these results is produced rapidly, however; at best the process is extremely slow, and it is not possible to judge how much is accomplished directly by abrasion and how much by mechanical weathering combined with deflation of the loosened particles. Disintegration and abrasion together produce fine material which is carried away by the wind. As long as Libya keeps its present climate the wind will remain the only important agent of erosion. Running water is a negligible factor, as shown by the total absence of modern stream channels in most of the region.

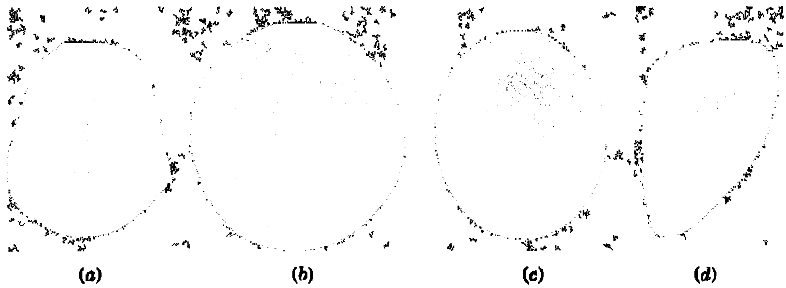


FIG 142 Comparison of stones fashioned by different agents of erosion. The long diameter of each stone is about 4 inches

a, Glaciated stone from till, New Haven, Connecticut.
b, Wave-worn beach cobble, Branford, Connecticut
c, Stream-worn cobble, Snake River, Washington.
d, Ventifact, Glass Mountains, western Texas.

Ventifacts. A conspicuous effect of abrasion in arid regions, and in a more limited way along sandy coasts in moist countries, is the polishing and peculiar shaping of pebbles that have lain for a long time on the ground in wind-swept areas. Sand driven by the wind grinds the pebbles smooth and slowly cuts upon many of them slightly curved facets that intersect along sharp edges (Fig. 142, *d*). If one side of a loose block of rock fronts for a long time toward the prevailing wind it finally develops into a facet that slopes upward away from the wind at an angle ranging from 30° to 60° from the horizontal. If the block becomes undermined so that it topples into a different position another facet develops in the same way. Thus the number and the shape of facets vary considerably; but a common form of sandblasted pebble is elongate, with three nearly equal faces that taper toward the ends. This form suggests a Brazil nut. Ordinarily the pebbles are made of quartz and of other minerals or rocks of superior hardness, and the facets have a high polish. To suggest the origin of the polish and the peculiar form, a wind-worn pebble is called a *ventifact* (*made by wind*).

Ventifacts are valued in geology not for their own sake, but because their presence in large numbers in a region where conditions now are not suitable for their formation indicates a radical change in the physical environment after their development. For example, an ancient gravel deposit widely distributed in northwestern Scotland contains numerous ventifacts, although the moist climate and consequent abundant vegetation do not permit any appreciable wind abrasion in that region at present. It is not to be assumed, however, that these wind-fashioned stones signify arid conditions at the time of their formation. Gravels deposited in New England near the end of the last glacial age contain abundant ventifacts. New England was barren and subject to wind abrasion at that time because the climate was cold and the ground was covered with sandy, stony debris released by the wasting glacier ice.

✓ *Wind Abrasion in the United States.* No part of the United States is comparable to Libya as a theater of wind erosion. Even the driest portions of Nevada, Arizona, and eastern California have rainfall enough for the development of stream channels, and it is clear that running water is the dominant agent of erosion and deposition in that region. Wind abrasion is conspicuous locally, as on the lower parts of the walls of narrow canyons through which sand-laden wind is forced with unusual violence (Fig. 143). Polished rock surfaces, usually grooved or fluted, are developed in this way. Areas floored with weak sandstone and exposed to exceptionally strong winds are abraded irregularly to form shallow basins that hold water after rains. The total effect of abrasion, however, is not large. If natural sandblasting operated widely in that region we should find polished and grooved surfaces widely distributed. Instead, most of the outcrops are rough from weathering, and many surfaces are dark with desert varnish, a peculiar shiny black coating of manganese and iron oxides which forms very slowly, partly through the activities of lichens and probably also by oxidation of mineral matter deposited when water held in pore spaces of the rock makes its way to the surface and evaporates.[1] Effective wind abrasion would prevent this slow accumulation on exposed surfaces and would remove all weathered rock particles from such surfaces.

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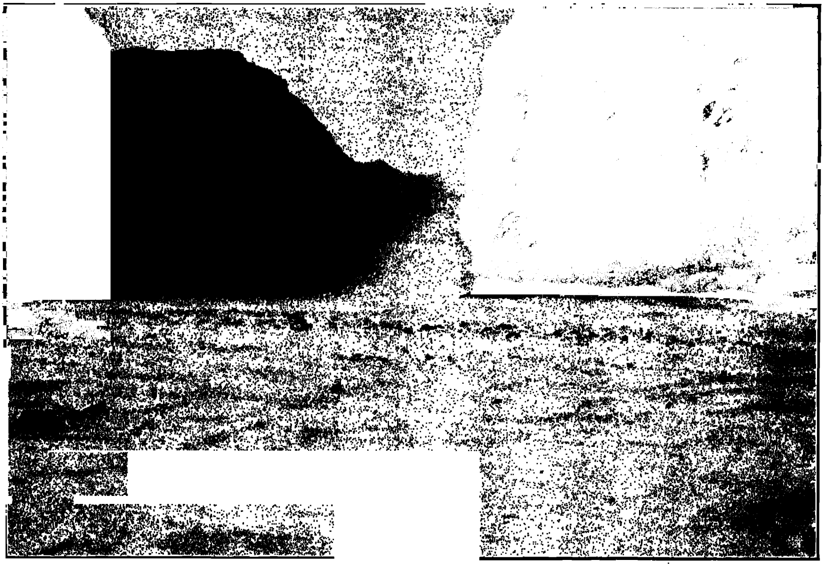
[1] Dark, shiny coatings of this kind are not restricted to arid regions. Excellent examples are found even along the Atlantic seaboard. However, the "varnish" is conspicuously displayed in arid lands only.

Pedestal rocks, which consist of wide caps supported by slender columns, are often cited as products of abrasion in the arid Southwest; but most of them result from differential weathering (p. 49). The winds help chiefly by removing loosened particles.



C. E. Erdmann.

FIG. 143. Cloud of sand raised by a moderate wind, head of Black Canyon, Colorado River, southern Nevada. Violent winds sweep sand through such narrow defiles with great force, polishing and grooving the lower parts of the rock walls.

It is concluded, therefore, that the only wind erosion of general importance in southwestern United States is performed by deflation of silt and sand (p. 202). Abrasion is a subordinate process, local in its operation.

WIND DEPOSITS

Like running water, wind loses its carrying power for various reasons and drops its load of debris. Some of the resulting deposits are only temporary; they are partially or wholly destroyed by the next strong wind. Other accumulations remain in place until by compaction and cementation they become sedimentary rocks (p. 260). Numerous layers of ancient sandstone are interpreted as *eolian* (wind-laid) deposits.

DUNES

Wind-blown debris, usually sand, accumulates to form rounded, elongate, or irregular hillocks known as *dunes*. The growth of a dune may be started by an obstacle, such as a stone, a bush, or an irregularity in the surface of the ground, which breaks the force of the wind; after the resulting heap of sand has grown to appreciable size it acts as its own windbreak and causes further deposition. Where sand is abundant and winds are strong some dunes grow as high as 100 feet or exceptionally, as in northern Africa, in southern Iran, and in the Great Dunes National Monument, Colorado, up to 500 or even 700 feet. If prevailing winds blow inland across a sandy shore a belt of dunes is formed (Fig. 144); such a belt, with some exceptionally large dunes, exists along the east coast of Lake Michigan, where the westerly winds blow across a large supply of beach sand. In arid regions, lack of vegetation and dryness of the ground permit the most effective accumulation of sand grains by the wind and therefore the most extensive development of dunes However, the popular idea that the entire surface of a desert is mantled with shifting sand is erroneous. Arabia is more widely mantled with dune sand than any other large region, and yet the sand covers no more than one-third of the total area About one-ninth of the Sahara is covered with drifting sand Large tracts in all deserts are floored either with bedrock or with loose stones. Fine-grained material is swept from these areas as fast as it forms; silt and clay particles are carried far away, and ordinarily the sand accumulates on the lower ground, particularly on the lee sides of low ridges where the force of the prevailing wind is broken, or at the windward bases of high, steep ranges across which the wind can not transport it.

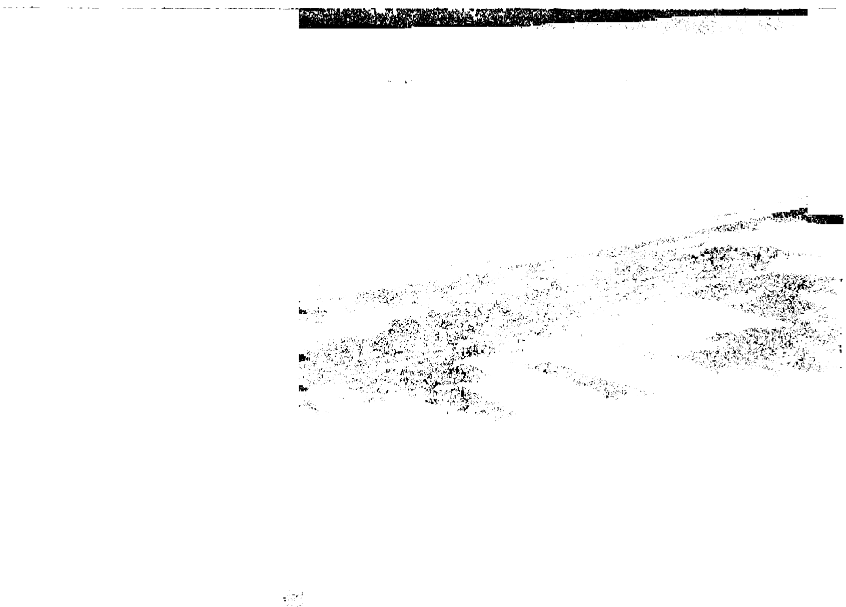
Even in arid districts the abundance of sand available for dune formation varies greatly with the character of the exposed bedrock In southern Nevada large areas in which the bedrock is chiefly limestone and shale are almost devoid of dunes. The most favorable bedrock is sandstone, which becomes loose sand as it disintegrates. In southern Utah and northern Arizona some thick formations of sandstone supply abundant sand to nourish dunes covering wide areas.

Forms of Dunes. Local conditions under which dunes are developed vary widely, and consequently there is a wide range in shape as well as size. A general distinction can be drawn between dunes resulting from fixed obstructions, such as hills, cliffs, and buildings, and those that exist independently of fixed surface features. Dunes in the latter class are capable of moving from place to place while retaining their characteristic forms. These free-moving dunes are common even under moist climatic conditions, wherever the supply of sand is so large that nearly all vegetation is either destroyed or prevented from gaining a foothold. However, in such regions both moisture and vegetation interfere to some extent with free movement of sand and thus cause many modifications in shapes of dunes. The most ideal forms of dunes, in their greatest scale of development, are found in extremely arid lands such as Arabia, the Libyan Desert, and southern Peru.

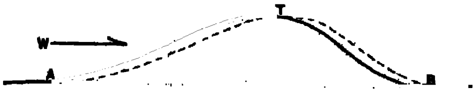
Free-moving dunes are of two fundamental kinds, with relation to the direction of the prevailing wind. These two classes are (1) *transverse* dunes and (2) *longitudinal* dunes. A wind with fairly constant direction blowing across an extensive source of loose sand (such as a sandy beach, Fig. 144) will build up long, parallel sand ridges that extend transverse to the wind direction. In cross-section such a dune has

SPENCE AIR PHOTOS.

Fig 144 Sand dunes migrating inland from Pismo Beach, California View looking northwest Note that the long, fairly regular dune ridges, transverse to the wind direction, tend to break up into shorter barchan-like dunes, seen in left foreground (See Fig. 147)

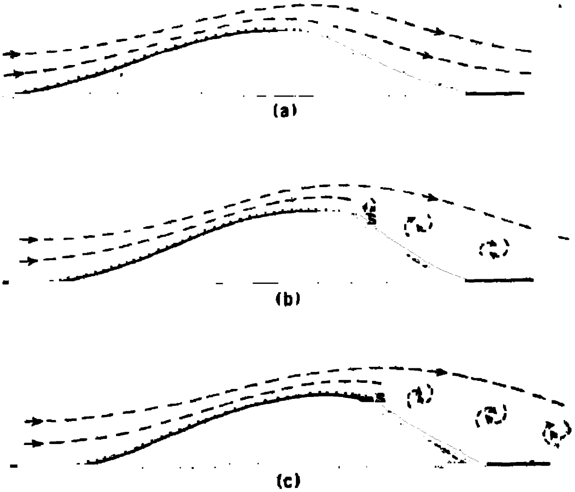


the form shown in Fig. 145. As the wind scours the windward slope, sand grains eroded from this slope are dumped over the crest, perhaps with other sand newly transported from the general source. Since most of this sand falls on the upper part of the lee slope, that slope continues to steepen until it reaches the *angle of repose*, which for dry sand ranges up to 34°. Further addition of sand then causes sliding, and as repeated slides occur the profile of the lee face (or *slip face*) of the dune becomes almost a straight line, with the steepness of slope con-



Modified from R. A. Bagnold.

FIG. 145. Generalized section through a transverse dune. Arrow shows wind direction. Broken line indicates that sand eroded from the windward side is added to the lee side of the dune.



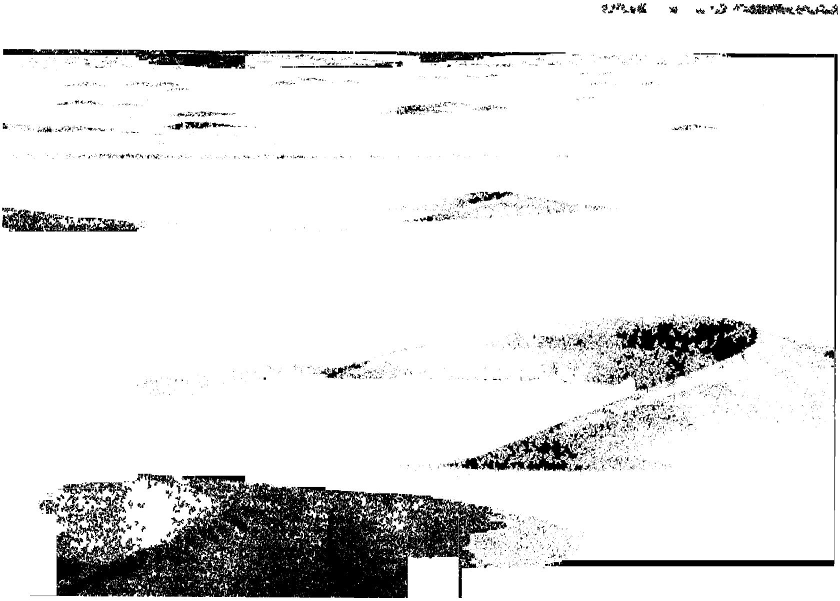
Modified from R. A. Bagnold.

FIG. 146. Development of *slip face* on a transverse dune. Up to a certain stage in growth of dune, wind flows evenly over crest, and surface of dune has a streamlined form, as in (*a*). (Dashed lines, with arrows, represent flow of wind.) Sand grains (represented by dots) are moved continuously across surface. As height of dune increases, flow of air is hindered on lee side, and eddies are formed there, as shown in (*b*). Sand then accumulates back of crest, at *s*, until slope becomes oversteep (broken line below *s*), when sliding occurs to a profile of equilibrium (full line below *s*). With repetitions of this action, the entire lee slope becomes a slip face, under control of gravity, as shown by full line below *s*, in (*c*).

trolled by gravity (Fig. 146). Thus the direction of the wind is clearly indicated by the steepness of the slip face in comparison with the gentler windward slope.

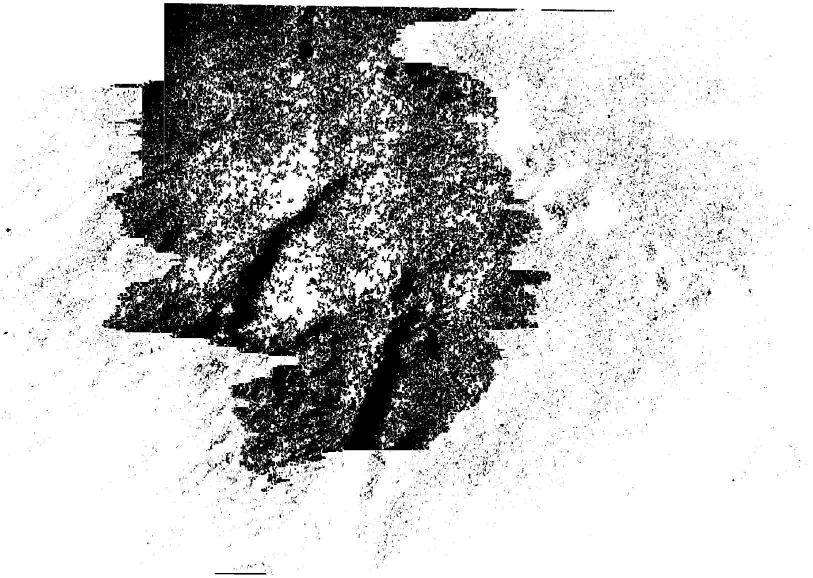
Transverse dune ridges are unstable, and with increasing distance from the source of sand a ridge tends to break up into individual hillocks. The wind continues to sweep sand not only over the crest of each hillock, but also along both flanks, building two long, curved tips pointing to leeward (Fig 147). Such a crescentic dune is called a *barchan* (bär'kăn). The barchan is the simplest, most symmetrical, and probably the commonest form of transverse dune. Any heap of sand that is freely exposed to the action of a steady wind will develop into a barchan. Thus the breaking up of a transverse dune ridge is only one method by which an array of barchans may be produced.

Like other transverse dunes, the barchan has a slip face and a gentler windward slope. Therefore the direction of the wind is indicated by



SHIPPEE-JOHNSON PERUVIAN EXPEDITION

Fig 147. Crescentic sand dunes (barchans) on the air route near Arequipa, Peru. Scale is given by the men, and by the airplane on the lee side of one dune. Note that the right (windward) side of each dune has a much gentler slope than the slip face

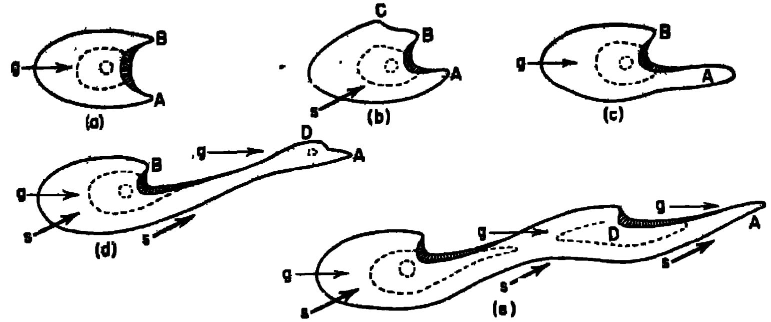


U. S. ARMY AIR CORPS.

Fig 148 Great longitudinal dunes in southwestern Algeria. Note the transverse dune ridges, especially conspicuous in foreground, which indicate a wind essentially parallel to main ridges and toward the observer. This direction is indicated also by the orientation of barchans. Main ridges probably are product of this prevailing wind combined with recurrent storm winds from upper left of view. (Shadow of airplane wing at right.)

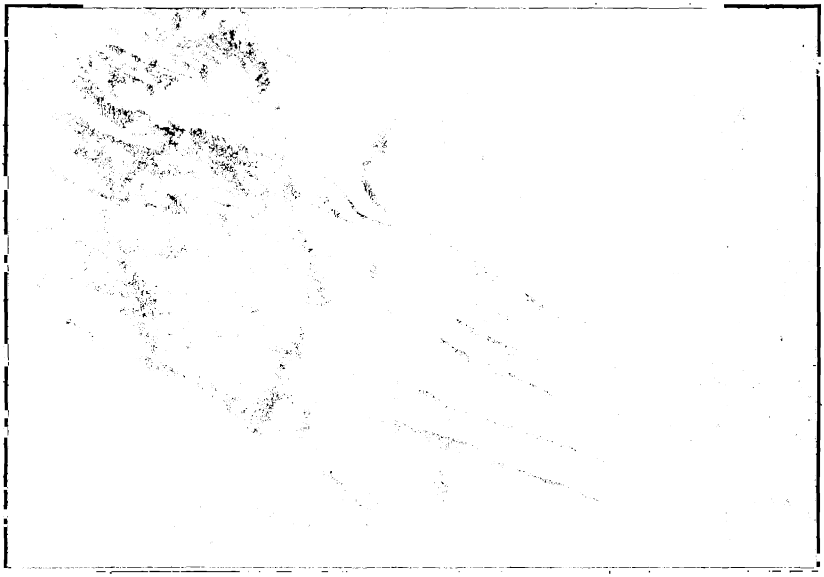
the contrast in the two faces, and also by the points of the crescent (Fig. 147).

Longitudinal dunes are distinct ridges, more or less regular, elongate in the general direction of prevailing wind. Over large areas of the African deserts, dunes of this type are remarkably developed, in a symmetrical pattern suggesting great windrows in a field of newly raked hay, with nearly uniform spacing (Fig. 148) Although the origin of such dunes is not yet fully understood, the evidence now in hand indicates strongly that two alternating directions of wind, making a considerable angle with each other, are responsible. Thus in Fig. 149, (*a*), a wind blowing most of the time from the direction *g* would form a typical barchan. Recurrent storms from the direction *s*, alternating



R. A. Bagnold, "The Physics of Blown Sand and Desert Dunes," William Morrow & Co., Inc.

Fig. 149. Possible development of a longitudinal dune by combined action of two wind directions. In (*a*), the prevailing wind *g* tends to form a simple barchan. A strong storm wind *s*, shown in (*b*), modifies the form of the barchan, displacing the crescent-point *A* and starting a new point *C*. Alternating action of the two wind directions produces the successive forms (*c*) to (*e*). Note that the plan of (*e*) is elongate generally in the direction of *g*, but has a somewhat staggered course. The ridge will also be very irregular in height and in many details of form. Compare the actual dune ridges of Fig. 148.



Spence Air Photos.

Fig. 150. Longitudinal dunes developed locally through the influence of irregular topography. A strong prevailing wind from the upper left sweeps sand through narrow passes between high hills made of bedrock. Long, narrow dunes, each tapering to a point downwind, result from concentration of the wind velocity along narrow lanes, with gradual loss of velocity on the lee side of the hills. Beyond the points of the longitudinal dunes, sand that is swept onward forms groups of small transverse dunes. Superstition Mountains, Imperial Valley, California.

with the prevailing wind *g*, would bring about the modifications shown in (*b*) through (*e*). Irregularities of pattern in the main ridges seen in Fig. 148 suggest a similar development, with a prevailing wind nearly parallel to the main ridges and recurrent storm winds from the left of the view. The spacing of the ridges and other features of these great longitudinal dunes present challenging problems in aerodynamics.

Longitudinal dunes of more limited extent are formed where strong winds sweep sand through valleys cut into a range of hills or into the edge of a plateau. The funneling effect sweeps the sand in a narrow path and deposits it in a nearly straight ridge that tapers to leeward (Fig. 150). Such dunes in Arizona and in southeastern California reach lengths up to 2 or 3 miles. At greater distances the funneling effect disappears, and sand carried beyond the tips of the straight sand ridges tends to form groups of small dunes.

Special Forms of Dunes. The growth of vegetation on sand-covered areas interferes with free movement of the sand and gives rise to many nondescript surface forms. At least one distinctive type of dune is developed under these conditions. Local absence of the protective vegetation permits the wind to start erosion of a "blowout" and to build the resulting sand into a *U-shaped* dune[1] around the borders of the depression. The general plan of a dune in this class may suggest superficially the form of a barchan. However, the two kinds of dunes differ fundamentally in origin and also in important elements of form. The tails or tips of U-shaped dunes point to *windward*—not leeward; and on the inside of the curve the slope is much gentler than on the outside (Fig. 151). Dunes of this kind occur in local groups in sand-covered areas of southwestern Kansas and in parts of the Navajo Indian Reservation in northern Arizona. Ancient curved dunes in Central Europe, now held firmly in place by vegetation, probably had a similar origin under harsh climatic conditions during the last glacial age (p. 189).

Abrupt irregularities of the ground give rise to complex and turbulent currents of wind, with resultant special forms of dunes that do not fit into any simple classification. For example, long, tapering ridges of sand are built on the lee sides of steep, narrow buttes; piles of sand accumulate against the windward slopes of high ridges; sand blown from a plateau surface over a cliff may build up a continuous slope at the base of the cliff. Dunes built under other special conditions are too numerous for an exhaustive treatment here.

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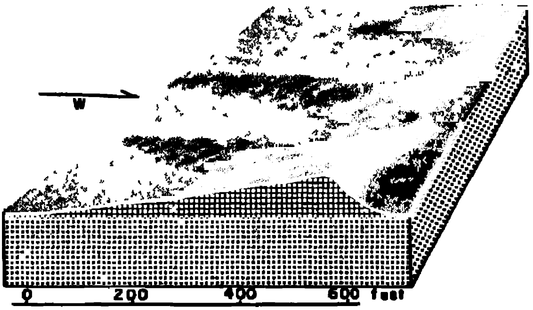
[1] Also called a *parabolic* dune. However, the common plan differs considerably from a parabola.

Migration of Dunes. Unless a dune is fixed in place, by either a topographic obstruction or a growth of vegetation, it is a shifting feature. If more loose sand is available than the wind can transport, the windward as well as the leeward slope of such a dune is built up, and the dune grows both in height and in the longitudinal dimension. Commonly, however, the windward slope itself is eroded by the wind, and the sand grains from it are moved over the crest, as explained in Figs. 145 and 146 (p. 211). By this subtraction of material from one side and addition to the other the dune moves forward slowly, provided the wind direction is fairly constant. In this way dunes formed along a shore migrate inland across low country (Fig. 144), and the movement halts only when grass or other vegetation gains sufficient foothold in the sand to hold it in place. In France and some other European countries belts of dunes moving from the coast have destroyed farm lands, forests, and villages. Along some parts of the Bay of Biscay and of the Baltic region the menace of dune migration has been ended by skillful planting of trees and shrubs. In the United States also considerable damage has been done by drifting sand, particularly near some of the Great Lakes and along many parts of the coasts, both East and West. It is necessary to build fences and other structures on the windward sides of railways and highways in some dune areas to check the velocity of the wind and thus cause the sand to be dropped where it will do no harm.



Modified from John T. Hack.

FIG. 151. Typical U-shaped dune, resulting from local wind erosion in deep sand, and deposition of the deflated sand around three sides of the resulting depression. Arrow shows direction of wind.

One of the largest belts of coastal dune sand is along the Bay of Biscay in western France; it is about 150 miles long and 2.5 to 6 miles wide. From this tract dunes travel inland at varying rates up to more than 100 feet per year, across a low swampy area. Numerous villages and considerable areas of farm land have been destroyed. At Lege the church was taken down at the end of the seventeenth century and built 2.5 miles farther inland to save it from the encroaching sand; 100 years later it had to be rebuilt again, because the sand had reached it once more.

Internal Structure of Dunes. In a dune that is growing on both windward and lee sides, successive layers accumulate essentially parallel to the surface slopes. A dune that is actively migrating has layers formed on the lee side only, and these are subject to rapid destruction as the dune advances. As shifting winds modify the earlier form, the layers are cut across, and their edges may be covered by new layers formed in different directions and at different angles. Therefore sand-dune deposits are characterized by extremely irregular *cross-lamination* (p. 271).

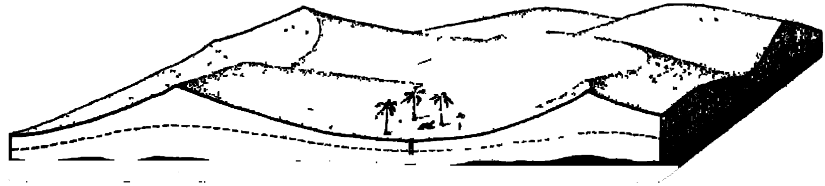


FIG. 152. A well in dune sand, northern Egypt. The dark portion of the block represents bedrock beneath the sand; the broken line shows the position of the water table.

Economic Value of Dunes. Although dunes generally are destructive and barren, in some arid countries they are the only natural reservoirs of fresh water. The sands absorb the scanty rainfall, the water table rises in a curve roughly parallel to the surface (Fig. 152), and at favorable localities shallow wells furnish enough water to form oases of considerable size; in northern Egypt large groves of palm trees are supported in this way. In some parts of western Texas, ranches get water from shallow wells put down at the very summits of large dunes.

Ancient Dunes. Over a wide area between Albany, New York, and the Adirondack Mountains there are numerous knolls that have the forms of typical dunes, although they are completely covered with grass and trees. Excavations made into many of these knolls reveal that they are composed of sand and finer particles deposited with the typical structure of dunes. It is thought that near the close of the last glacial age, when much of the surface was mantled with deposits washed from the wasting ice and there was little protecting vegetation, the wind swept great quantities of rock debris from the Adirondack region and formed dunes on the adjacent plain. Similar features are common in central Connecticut, Ohio, Indiana, and other glaciated areas.

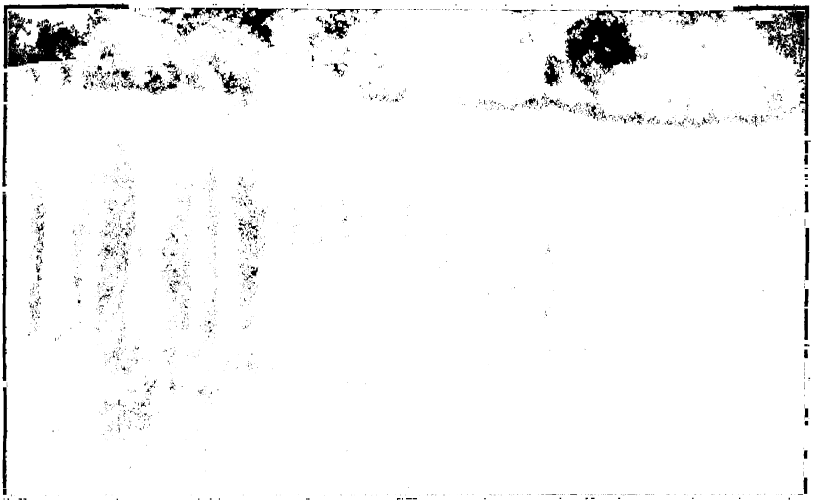
In Bermuda many artificial cuts reveal a peculiar limestone that has the cross-laminated structure of dunes. Evidently this is an old dune deposit similar to that now forming along the Bermuda shores, where the sand consists of coral limestone ground up by wave action. Dunes are built of this sand, the grains become cemented, and the result is new limestone like that exposed in the cuts. Under exceptional conditions various other minerals and rocks have furnished debris that has been built into dunes. An area of 500 square miles in New Mexico is covered with dunes made of snow-white gypsum which was derived from disintegrating beds of gypsum exposed at the surface. Some of this gypsum sand is still shifting with the winds; but in many of the older dunes the grains have become firmly cemented. This large area has been set aside as The White Sands National Monument.

The examples cited above relate to dunes which, though they are no longer active, were formed in fairly recent geologic time. In the older sedimentary strata we recognize dune structure, now preserved in firmly cemented sandstones, which gives a record of shifting sands in ancient geologic periods.

LOESS

What has become of the great quantities of fine material removed from land surfaces by deflation? Part of the answer is given by a peculiar yellowish, fine-grained sediment that covers vast areas in Asia, Europe, and North and South America. Typically it has no horizontal stratification, like that in ordinary sedimentary formations, but occurs in a single massive layer, 20, 50, or even more than 100 feet thick. On the other hand, it is cut by nearly vertical surfaces that divide the deposit into rough columns; for this reason it has the remarkable property of forming high bluffs along valley sides in spite of its soft, earthy character (Fig. 153). This sediment, so similar in widely separated continents, is known by the German name *loess* (lûs).

Although loess is exceedingly fine grained, examination with a powerful microscope reveals that a large proportion of the material is not decomposed but consists of fresh, sharp-cornered particles of feldspar, quartz, calcite, mica, and numerous other minerals mingled with clay It is evident, therefore, that much of the material was ground up mechanically, and that the particles thus formed were not affected by chemical weathering before their deposition. Shells of land snails and bones of land animals are found in the deposits. Moreover, loess forms a blanket of variable thickness, covering older hills and valleys of very irregular surfaces. The wind is the only known agent that could deposit in this way sediments that are uniformly fine grained. General lack of stratification is to be expected in wind-laid silt, since the deposit at any time is irregular, and after deposition it is worked over with the underlying sediments by rain, frost, worms, and growing plants. Slender vertical tubes that are common in loess appear to represent the stems and roots of successive generations of plants that were buried by the accumulating sediment.



U. S. Geological Survey.

FIG. 153. Typical bluff of loess with vertical columnar structure. Near Beverly, Missouri.

Locally the unstratified loess grades laterally into deposits of similar silt that are stratified and have other characteristics of water-laid strata. Such deposits represent the action of running water on slopes and in local channels, both while the wind was building up the more characteristic accumulations of loess and during the time that has elapsed since the original deposition.

In central Europe, in the Mississippi Valley, and in eastern Oregon and Washington, the fine material forming the loess probably was supplied by the dried floors of temporary lakes and by floodplains of streams that drained from the wasting glaciers during the glacial ages (Chap. 9). While these wide sheets of sediments lay unprotected by vegetation the wind picked up the finest-grained material and spread it out on the uplands; the abundance of fresh rock flour ground up by the ice movement explains the large percentage of unweathered, angular particles derived from numerous kinds of rocks and minerals.

In northern China an area as large as France is covered with loess having a maximum thickness of several hundred feet. Streams have cut canyons into it, and the quantities of yellow silt continuously eroded from the area are responsible for the names of the Yellow River and the Yellow Sea. Without reasonable doubt this vast deposit of loess represents the agelong accumulation of fine debris blown from the Ordos and Gobi deserts and from other parts of the arid interior of China. Thus, although the deposit that formed this loess was quite different in its origin from that in Europe and the United States, the material is similar because it also was prepared mechanically; in the deserts of Asia it is the result of disintegration and wind abrasion in an arid climate, and therefore fresh, angular particles make up a good proportion of the deposit. The Chinese loess forms the richest soil of China, and for centuries many of the humbler farmers in the hilly districts have made their homes in artificial caves dug into the steep bluffs. In these districts some of the roads used for a long time have become narrow steep-walled canyons as the loess loosened by travel has been removed steadily by wind and rainwash.

Loess and related deposits represent only the most conspicuous accumulations of wind-blown silt; unquestionably, dust carried in the upper air, supplied either by deflation or by eruptions of fine volcanic ash (Chap. 14), settles slowly on the entire land surface and on the sea floor. Probably almost every layer of sedimentary rock contains at least a small amount of eolian deposit. Conceivably some thick-bedded ancient siltstones may consist of loess laid down in various periods of geologic history and preserved by favorable circumstances to become part of the sedimentary record. Most loess deposits, however, because of their elevated positions on the continents and the weak character of the material, are subject to rapid destruction by erosion.

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An excellent description and explanation of features in the world's most famous arid region.

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Has good descriptions and explanations of eolian features, with excellent halftone illustrations.

9. Deserts on the March; by P. B. Sears. 231 pages. Oklahoma University Press, Norman, 1935.

A well-written book, in popular style, dealing with the destruction of the soil begun by man's activities. Conditions in the "Dust Bowl" are explained.

CHAPTER 11

MARINE EROSION AND DEPOSITION

The gradual crumbling of the land under the pounding of the surf is a fact familiar to all who have lived near the shore, especially along coasts composed of weak rocks. The seaward shores of Cape Cod and the islands near it yield to the sea from 1 to 6 feet each year, and at the present rate these lands will disappear entirely at the end of about 4000 years, leaving only submarine banks to mark their former positions. Certain stretches of the Yorkshire coast in England have been worn back a mile since the Norman conquest, and 2 miles since the time of the Romans. In the space of 200 years the town of Egmont, on the Dutch coast, was undermined and entirely destroyed by the persistent work of the waves. Cliffs near Dover on the English Channel are receding at the rate of 15 feet each year, and east of them the Goodwin Sands, now a shallow submarine bank, were formerly an island. The island of Helgoland off the mouth of the Elbe, once very large, was being eaten away so rapidly that it would by this time have disappeared entirely had erosion not been checked, near the end of the nineteenth century, by the construction of a strong sea wall.

All these examples concern coasts formed of weak materials, such as chalk, clay, and sand, but the effect of waves on resistant rocks is the same except that erosion takes place much more slowly.

A study of shore processes has practical value, for destruction of the land by the sea can be stopped for long periods, measured in human terms, as is demonstrated by the case of Helgoland. Thousands of thickly populated districts where shore property is valuable have been artificially protected in this way. This study also furnishes answers to such broad questions as these: How far can the destruction of land by the sea be carried? Why do not beaches, bars, and other wave-built features stop marine erosion? And above all, since the sea-covered area of the Earth is three times as extensive as the land area, and since the Earth is admittedly very old, why have the continents not already been destroyed by marine erosion?

Functions of the Sea. The sea, covering 71 per cent of the Earth's surface, is a direct and energetic agent of erosion. Bit by bit it tears away the margins of the land and thereby aids the streams in their work of wearing down the continents. Its floor is the final resting place of all the rock waste eroded from the lands by waves and currents, and by streams, subsurface water, glaciers, and the wind. This land-derived sediment is shifted for a while along the sea floor and, coming to rest, is ultimately solidified into the sedimentary rocks that form a conspicuous part of the Earth's crust.

Indirectly, too, the sea exercises a great influence on the development of the Earth's surface. Throughout the hundreds of millions of years of geologic time the general temperature of the Earth has never fallen below the freezing point or exceeded the boiling point of water. For this stability the nearly uniform rate of radiation of energy by the Sun is chiefly responsible, but the sea also serves as an important stabilizer by storing up excess heat against times of lessened solar radiation. The great currents that carry warm water from the equatorial zone into the higher latitudes and the currents that transfer cold polar water toward the equator likewise exercise an important influence in distributing heat more evenly upon the Earth's surface and in decreasing the contrasts between climatic zones. Moreover, in the last analysis, evaporation from the surface of the sea supplies all the moisture borne by the winds to fall upon the lands as rain and snow. Thus the sculpture of the lands by streams, glaciers, and subsurface water is ultimately dependent on the great reservoir of the sea.

Composition of Sea Water. About 3.5 per cent by weight of sea water consists of dissolved mineral substances, most of which have come from the lands. When sea water is evaporated, more than three-quarters of its mineral content is precipitated as common salt ($NaCl$). Among the many other substances present are calcium carbonate ($CaCO_3$) and silica (SiO_2). Both these substances are removed by chemical precipitation and by marine organisms which use them in forming their shells. The total quantity of the salts is astonishing. If precipitated and crystallized into a bed of solid salts, it would form a layer about 125 feet thick over the entire surface of the Earth.

Terminology. For the purpose of our discussion we shall define the sea as the entire body of confluent salt water of the globe. This leaves out only the separate inland bodies of water which, whether fresh or saline, belong in the category of lakes (Chap. 8). *Sealevel* is the surface of the sea, and according to variations of depth we have deep sea and shallow sea, each with its characteristic processes, deposits, and living inhabitants. Shallow seas that lie upon a continent and are nearly landlocked are *epeiric seas* (Greek *epeiros*, a continent). At present there are but two good examples, the Baltic Sea and Hudson Bay, but at times in the past, when the continents were far more widely flooded than now, epeiric seas were of vast extent and of great importance. In fact, most of the sedimentary rocks of the present lands were formed in seas of this kind.

Movements of Marine Water

Sealevel; Tides. Although it is customary to use sealevel as a common datum of reference, the surface of the sea is not a perfect sphere. Its polar diameter is about 27 miles less than its equatorial diameter, and in addition there are irregular local departures from the spherical surface, the most obvious of which is the periodic distortion imparted to it by the tides.

The combined gravitative attraction of the Moon and the Sun pulls the surface of the sea into two low but vast bulges, one on each side of the Earth. These bulges cause the tides. They are fixed with respect to a line connecting the Earth with the Moon, but, since the Earth in its daily rotation turns from west to east, they seem to move around the Earth from east to west. Far out at sea, the surface merely rises a little and then subsides again as each tidal bulge passes; but where the bulge impinges against a coast the water is dragged forward, piling up on the shore and then receding.

The height of tide is determined largely by the configuration of the coast. On open, exposed coasts it is not more than 6 or 8 feet, and in nearly landlocked embayments such as the Gulf of Mexico it is only 1 or 2 feet. In estuaries that open out toward the advancing tide, however, the water piles up as it is crowded forward into the ever-narrowing bays. This brings about exceptional conditions, such as those in the Bay of Fundy, where the tide rises normally 30 to 40 feet and exceptionally as much as 50 feet.

These immense tidal bodies of water, ceaselessly moving in and out of bays and estuaries and along coasts, set up strong currents which scour the bottom and transport quantities of sediment. The rise and fall of the tides also aid indirectly in the attack of the waves on the land by increasing their vertical range.

Ocean Currents. The sea is affected by broad drifting movements confined to the several oceans. These are *ocean currents.* They are kept in operation chiefly by winds, such as the Trades, that blow steadily in one direction, and secondarily through the movements set up by differences in density of the water from region to region.

Many ocean currents are no more than faint drifts. Few travel more than 10 or 12 miles in 24 hours through open water. They affect the sea floor on the continental shelves and continental slopes and may pass over it with sufficient velocity to move sediment. In addition they exert a strong indirect effect by greatly modifying the climates of the lands they pass to windward. The equatorial currents carry warmth to the polar seas, and polar currents bring low temperatures and icebergs into middle latitudes.

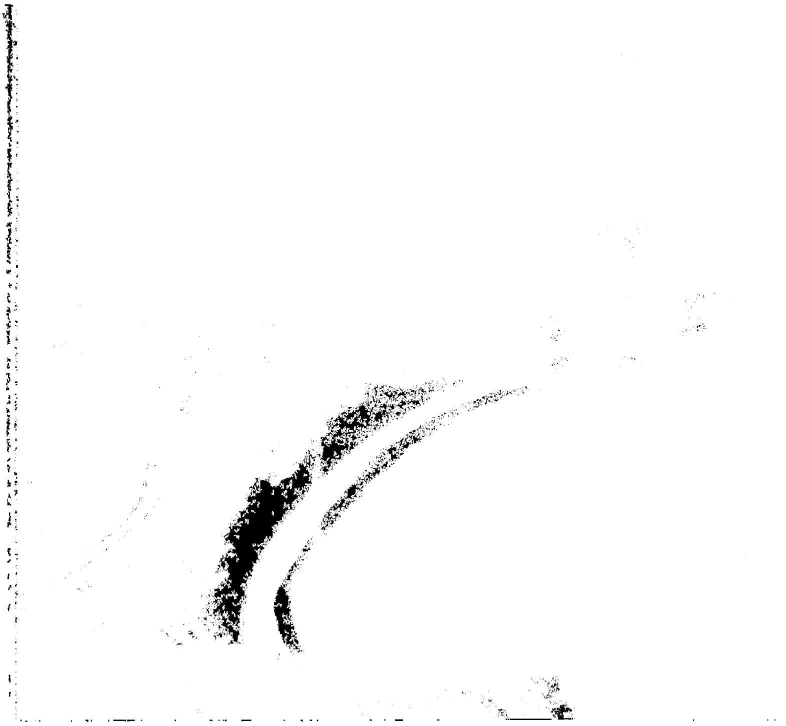
Waves and Rip Currents. Waves are generated by the wind blowing along the surface of the sea, which is thereby thrown into undulations. Once formed, these undulations are maintained and increased by the pressure against their windward sides and so are driven forward in endless succession. The wave form travels forward somewhat like the rippling of wind across a field of grain, making the stalks bow as each wind wave passes, but allowing each to return to its former position. However, the friction of the wind that generates waves also causes slow motion of the water in the direction of wave travel.

The heights (of crests above troughs) and lengths (from crest to crest) of waves increase with the velocity of the wind, with its duration in a given direction, and with the expanse of open water to windward. For this reason small bodies of water and protected embayments of the coast are not affected by great waves.

Wave motion decreases so rapidly with increasing depth of water that not even great storm waves can disturb the bottom at depths greater than a few hundred feet. Sea floors off exposed coasts are generally affected by waves to a depth of 200 to 300 feet, and exceptional storm waves move fine sediment at depths as great as 600 feet.

When a wave passes into shallow water, it becomes higher and shorter and its front steeper and more deeply concave until the crest arches forward, loses its support, and collapses in a rush of water, forming a breaker (Fig. 158). The water in the breaker, unlike that of the unbroken wave, moves forward rapidly. The height of the wave determines the depth at which it will break, and therefore small waves break in the very shallow water close inshore, whereas great storm waves usually break farther out, where the water is 10 to 20 feet deep.

Where large breakers move landward in shallow water the cumulative landward movement of the surface water is compensated by *rip currents*—seaward-moving lanes of water, irregularly spaced, flowing through the zones of breakers and dissipating in the deeper water beyond. At times such currents are dangerous to surf bathers.



SPENCE AIR PHOTOS.

Fig. 154. Coast near Laguna Beach, California, showing wave-cut cliffs and shallow bays fringed with beaches. The tide is low; hence the wave-cut benches seaward of the cliffs are exposed.

Longshore Currents. Longshore currents are set up by waves that strike the shore obliquely and thus have a component of motion along the shore. Because nearly all waves are oblique to the shore, longshore currents are very common. Most longshore currents are fairly steady because they are caused by waves that are generated by prevailing winds. Sweeping rock waste along, the currents tend to remove it from exposed headlands and to deposit it in protective bays (Fig. 154). They are largely responsible for the formation of beaches, spits, and bars. The fine sand of Daytona Beach, Florida, includes minerals that occur, not in the rocks of the Florida coast but in the rocks of the Georgia and Carolina uplands. For this reason the sand is believed to have been swept alongshore from farther north, where the rivers of Georgia and the Carolinas bring it down to the sea out of the southern Appalachians.

THE ATTACK ON THE SHORE

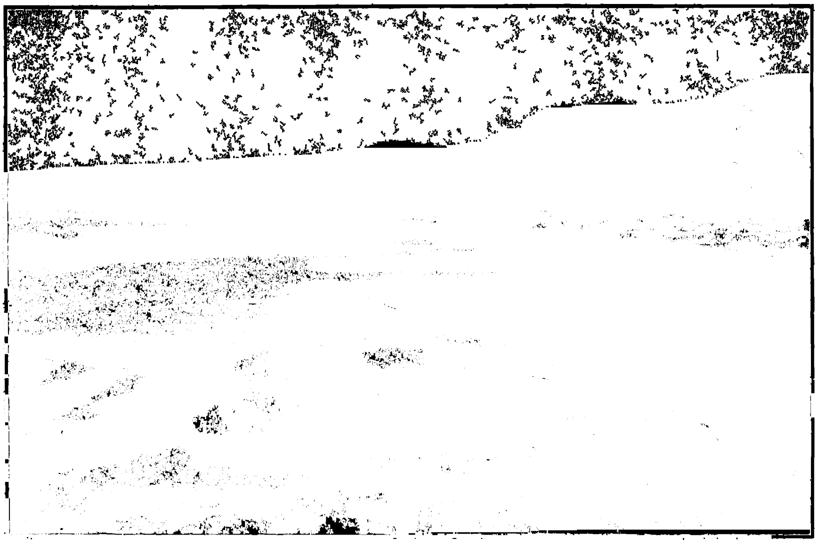
The sea, like other agencies discussed hitherto, uses the energy inherent in its motions in performing work. This work, like that of streams, can be classified as erosion (hydraulic action, abrasion, solution, and transport) and deposition. Let us consider, in turn, each of these processes (except solution, the results of which are not conspicuous in most places).

Hydraulic Action. The damage done by storms to piers, breakwaters, and similar structures is a rough yardstick for measuring the force of sea waves. At Wick, Scotland, during a great storm in 1872, a solid mass of stone, iron, and concrete weighing 1350 tons was torn from its place at the end of a breakwater and dropped unbroken inside the pier. The damage was repaired with a much larger block, weighing 2600 tons, but, in a storm five years later, the waves carried away this one too. It is not surprising, therefore, that the faces of cliffs are shattered by the impacts of storm waves. Hydraulically compressed air is an additional factor important in eroding cliffs of fissured and jointed rocks. The air in the fissures, violently compressed by the impacts of the waves, acts as a wedge, springing out large pieces of rock, especially on steep, cliffed coasts.

Abrasion. Abrasion is another important factor in the erosion of coasts by the sea. In shallow water the waves are well equipped for abrasion because they are supplied not only with the debris which they themselves wrest from the land, but also with waste contributed by streams. Sand and pebbles are rubbed together and ground against the shore by every breaking wave, and the erosive process is enormously hastened by the great waves of occasional storms, which fling stones and even boulders against the land, and then return to the attack with new weapons in the form of fragments of the eroded shore itself. By this process the sea slowly advances on the land as the opposing cliffs yield ground. The rock fragments dislodged from the cliffs are angular when first seized by the waves, but as they are handled by the sea they become smoothed and rounded (Fig. 142, *b*) by abrasion, until they come to resemble stream-worn stones.

Transport and Deposition. The water piled up against the shore by breaking waves is returned by seaward-moving currents, which carry rock particles with them. The rush of breaking waves up the beach carries more water to the beach than the corresponding backwash carries from it, because some of the water sinks into the beach.

In addition, the onrush has greater velocity than the backwash. The result is a size sorting of the rock fragments being handled; the sand and gravel tend to accumulate on the beach, while the silt and clay are gradually moved seaward. The larger fragments are dragged back and forth near the shore, and as they are abraded to smaller and smaller sizes (becoming rounded or flattened during the process) they are gradually shifted seaward. As they grow fine, they are moved in suspension rather than by traction, and at length each particle drops out in places too deep to be affected by motion of the water.



R F Flint, Geological Society of America.

FIG. 155. Wave-cut cliff and wide wave-cut bench cut across steeply inclined sedimentary rocks north of Bonne Bay, Newfoundland, exposed at low tide. The beach is unusually scanty, it consists of a few rounded stones and boulders.

The Shore Profile.[1] As waves dash against the land they carve a *wave-cut cliff* (Figs. 154, 155, 156) which slowly retreats under their attack, in some places at rates of several feet per year. The top of the cliff is undermined, falls bit by bit to the cliff base, and forms a *beach* made of wave-handled debris worn from the cliff and therefore consisting of sand, pebbles, or coarser fragments, according to the kind of rock into which the cliff is cut. The beach is added to by debris brought by the waves from elsewhere along the shore; indeed, some beaches have been made chiefly in this way. Sometimes the beach is partly cut away, sometimes it is greatly added to, but always it consists of rock waste in intermittent transit from the cliff to the deeper water offshore. If the rock forming the cliff is cut by joints and fissures, the waves commonly widen them and in places quarry out blocks, leaving great masses of rock standing isolated.

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[1] Technically, the *shoreline* is the line along which the sea intersects the land, the *shore* is the narrow zone lying between the low-tide shoreline and the high-tide shoreline, and the *coast* is a strip of indefinite width extending landward from the shore.

As the wave-cut cliff slowly retreats under the attack of the sea, it leaves behind it an ever-widening rock platform. This is the *wave-cut bench* (Figs. 154, 155, 156). Its surface, submerged at high tide and emerged at low tide, is gradually lowered by abrasion as waves and seaward-moving currents sweep debris back and forth across it. The bench is usually covered with gravel or sand, the net travel of which is slowly seaward. This sediment finally lodges in the deeper water at the outer edge of the bench, forming a *wave-built terrace* (Fig. 156).

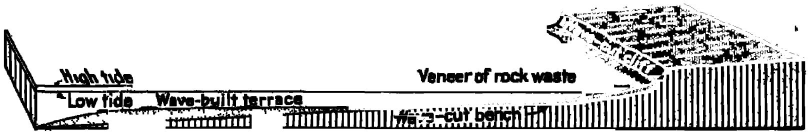


FIG. 156. Wave-cut cliff with beach at its base, wave-cut bench veneered with rock waste, and wave-built terrace made of accumulated debris. The exact form of the wave-built terrace is conjectural.

The combined profiles of the cliff, the wave-cut bench, and the wave-built terrace together form the *shore profile* which, under the continued influence of the waves, gradually changes its form. The cliff retreats, the wave-cut bench is widened landward by erosion, and the wave-built terrace is widened seaward by deposition.

The Graded Profile. The power of waves beating against a coast bordered by deep water immediately offshore is applied entirely to eroding the cliff, because the waves lose no energy by dragging bottom before they reach the shore itself. As the cliff retreats and the wave-cut bench is correspondingly widened and the cliff debris increases in volume, more and more wave energy is spent in grinding up rock fragments and in transporting the finer particles to deep water. Thus the work to be done increases, while on the contrary the power of the waves to do work along the shore steadily diminishes. Therefore a time comes when these two factors are balanced; when, in other words, there is a close adjustment between the amount of rock waste on hand and the amount of wave energy available to deal with it. This delicate adjustment—a mean through which the opposed forces continually fluctuate—is recorded by the steepness of the curving shore profile (Fig. 156), which, by slow degrees, becomes a *graded profile,* analogous to the graded profile of a stream (p. 81). The inclination of the graded profile depends chiefly on the strength of the waves and the kind of rock waste being handled.

Any disturbance of this balance results in adjustment that tends to restore balance. An increase in the supply of debris, for example, makes the profile less steep and slows up erosion of the cliff until the increase has been got rid of by seaward transport; an increase of wave energy steepens the profile and speeds up the attack on the cliff until the resulting increased supply of rock waste again becomes adjusted to wave energy. The maintenance of the graded profile is illustrated by the simultaneous retreat of three wave-cut cliffs of silt and clay along the shore at La Jolla, California. During the 12-year period 1918–1930, cliff *A,* 21 feet high, was cut back 20 feet into the land. During the same period cliff *B,* 33 feet high, receded 15 feet, while cliff *C,* 54 feet high, yielded only 10 feet. Expressed in terms of equilibrium, this means that, for each foot of retreat, the highest cliff yielded the greatest volume of waste, and this had to be handled and distributed by the waves before the cliff could be further eroded.

The evolution of a new shore profile into a graded profile is like the evolution of part of the long profile of a very young stream into the graded profile of a mature stream. Like the graded long profile of a stream, the graded shore profile flattens out as erosion and deposition continue.

DEPOSITIONAL SHORE FEATURES

The beach and the wave-built terrace have been mentioned; these are primary depositional features. In addition to these many shores are also provided with spits and bars.

Spits and Bars. Longshore currents sweep debris from headlands into the deeper water of bays, where, losing velocity, they dump it in embankments somewhat resembling railroad fills. As long as their far ends are free, these embankments are termed *spits* (Fig. 157). Waves drag bottom on the embankments and build them above sealevel. Spits not seriously impeded by transverse currents continue to be built across the bays, closing or nearly closing them, and thus form *bay bars* (Fig.

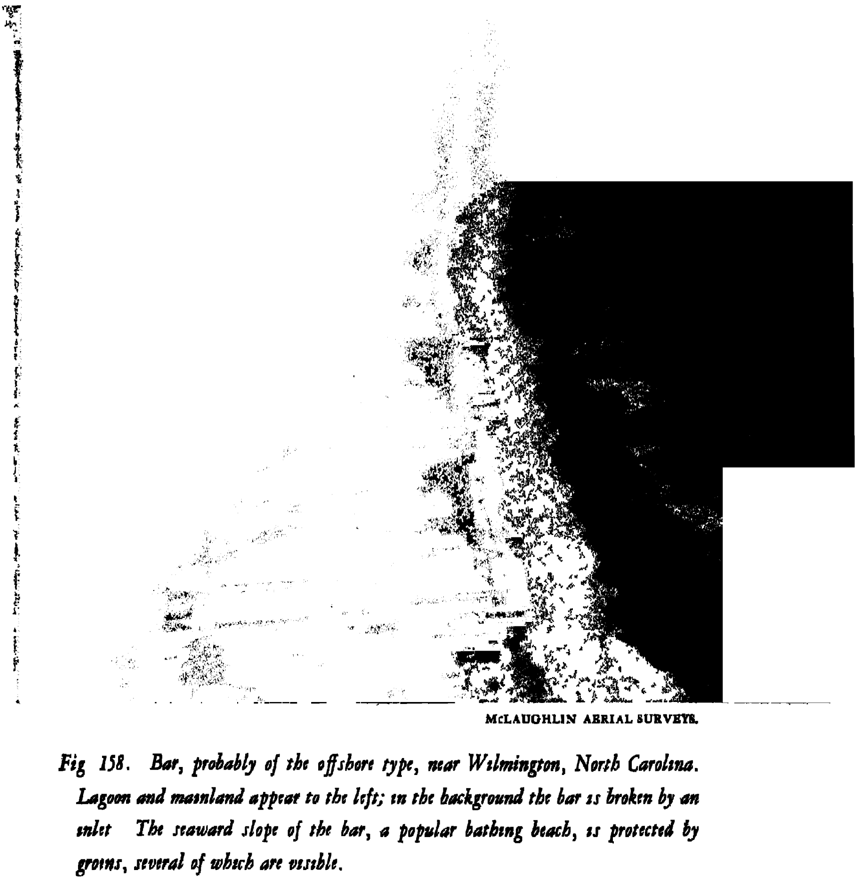
FAIRCHILD AERIAL SURVEYS

Fig. 157. Poponesset "Beach," a spit evolving into a bay bar along an embayed coast. Cotuit, Cape Cod, Massachusetts. The free end of the spit is being turned shoreward by waves and currents moving past it.

162). Other bars are built in the sheltered water between islands and the mainland, "tying" the islands to the mainland.

As a result of abundant stream deposition from the land, or warping of the crust, or other causes, the offshore slope of the bottom may be less steep than the slope required for a graded profile. Under these conditions the line of breakers lies well offshore. The waves drag the shallow bottom and pick up sediment which they carry forward to the line of breakers. Here they drop it, building up a narrow ridge or *offshore bar* (often popularly called a "barrier beach") parallel with the general trend of the shoreline (Fig. 158). Wave excavation continues to seaward of the bar, until the graded profile is attained. The bar itself may be built above sealevel by storm waves, inclosing a shallow lagoon [1] between it and the mainland. The width of the lagoon depends on the seaward slope of the bottom and the size of the storm waves. It is commonly a mile or more and may be several miles. The ideal offshore bar is thus a wave-built feature, in contrast with the ideal bay

[1] Large lagoons are also called sounds but not all sounds are lagoons. A sound is merely a wide strait.



McLAUGHLIN AERIAL SURVEYS.

Fig 158. Bar, probably of the offshore type, near Wilmington, North Carolina. Lagoon and mainland appear to the left; in the background the bar is broken by an inlet The seaward slope of the bar, a popular bathing beach, is protected by groins, several of which are visible.

bar, which is built by longshore currents. Doubtless most bars have been built by interplay of the two processes.

In general, the Atlantic and Gulf coasts of the United States exhibit nearly continuous bars, with bays and lagoons developed on a large scale. At Cape Hatteras (Fig. 159) the bar is about 20 miles from the mainland, but along the east coast of Florida, as at Palm Beach and Miami Beach, it is near shore, and the lagoon in consequence is very narrow. Both Atlantic City, New Jersey, and Galveston, Texas, are built on bars. In 1900 a disastrous flood occurred at Galveston when high seas, driven by a hurricane, rose 15 feet above their normal height.

In the graded-profile concept, we can now see the answer to the question often asked, and repeated at the beginning of this chapter: Why do not beaches and bars stop erosion by waves? Like their counterparts, the bars built by streams, the spits, bars, and beaches built by sea waves and currents are temporary, shifting structures. Just as a single flood may modify or destroy one of these structures or build a new one, so a single storm may work great changes along a shore. The cliff and bench are far more durable features.

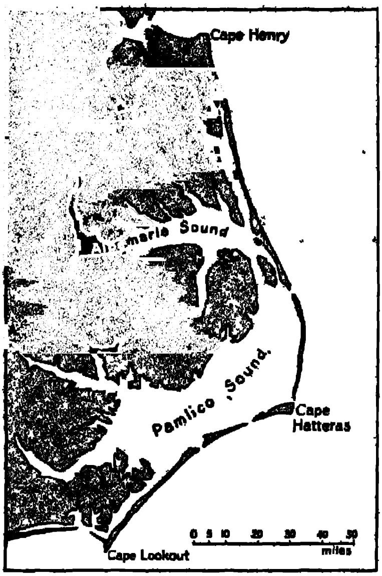


FIG. 159 Large bars inclosing bays and lagoons, in the region of the Carolina capes.

SHORE PROTECTION

Buildings and other costly structures standing close to the shoreline frequently require protection from the erosional inroads of the sea. Three general methods are employed to combat erosion of the shore in such places. One is the construction of seawalls along the shore and of breakwaters offshore. The waves are obliged to attack and destroy these structures before they can resume their erosion of the ground back of them. A second method is the construction of groins—low walls extending from the shore out into the water (Fig. 158). Groins are effective in many places where longshore currents are loaded with debris, because they check the currents and cause the deposition of the debris between the groins, thus forming a new beach which protects the land behind it. A third method is the hauling or pumping of sand from elsewhere on to the strip of shore to be protected. Ordinarily this operation has to be repeated annually and is used commonly in conjunction with the construction of groins. As shore property becomes more and more intensively utilized, protection of the shore will become an increasingly important problem.

THE GEOMORPHIC CYCLE ALONG COASTS

The chief erosional shore features—the wave-cut cliff and bench—occur in combination with the depositional forms—the wave-built terrace, the beach, and the various kinds of bars—giving seemingly endless variety to shorelines. Yet upon analysis these combinations are found to be the result of orderly and systematic development conditioned principally by the initial configuration of the coast before the shore processes began to sculpture it. In order to visualize this sequence, let us select two ideal initial coasts of very different types and trace the changes that will gradually take place in each, grouping them into stages of youth, maturity, and old age, as was done in the discussion of the fluvial cycle.

Bold Embayed Coasts. A bold coast marked by deep bays between pronounced headlands, with comparatively deep water and some islands offshore, is shown in Fig. 160. This general type of coast occurs in northeastern North America and most of western Europe. Its irregular shoreline is the result of depression of the land or actual rise of sea-level, or both, drowning the valleys and thus forming the bays. Wave attack is concentrated on headlands and islands, as the waves can not effectively reach the sheltered bays (Fig. 161). Cliffs are cut into the seaward shores of the islands and exposed headlands, and those parts of the headlands where the rock is weakest are quickly indented, forming little coves between small promontories of more resistant rock. As the headlands are cliffed, the debris first eroded sinks out of sight into the deep water at their bases. Gradually, however, the headlands are cut back until the rock waste supplied by their erosion is built up into the zone of wave action, where it forms beaches fringing the cliffs Longshore currents generated by oblique waves sweep past the cliffed headlands and carry part of the beach material into the deeper water of the bays where it forms spits and bars. Figure 162 shows several spits, a bay bar, and an island tied to the mainland.

While these features are being built with the debris of the crumbling headlands, deltas are being built out by the larger streams into the heads of some of the bays, which become fringed with tidal marshes (p. 155) in protected stretches where aquatic vegetation can grow undisturbed.

As the islands are cut away, the headlands cut back, and the bays correspondingly shortened, the shoreline becomes much simplified (Fig. 163). The truncated headlands are now connected by nearly continuous bars, and the bay heads are filled with sediment washed in by the streams. This simplification of the shoreline is the work of the stage of youth.

When the headlands have been completely cut away and the bay-head fillings removed, the shoreline is said to be mature (Fig. 164).

The waves can now attack the entire shoreline, except that stretches composed of weak rocks will continue to form indentations, while those composed of stronger rocks will form promontories. Along the whole shore are cliffs as high as the height of the land above the sea. The submerged wave-cut bench is as broad as the original headlands were long, a wave-built terrace extends seaward beyond it, and the combined profile of the two features has become a graded profile, across which waste from the cliffs is slowly shifted, adding to the terrace and completing the filling of the bays with sediment. A long period ensues, marked only by retreat of the cliffs and widening of the bench and

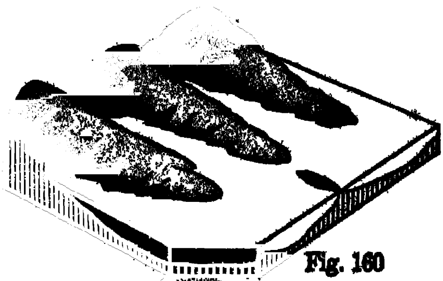


Fig. 160

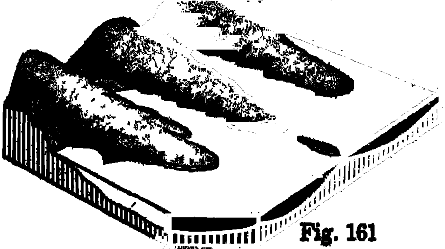


Fig. 161

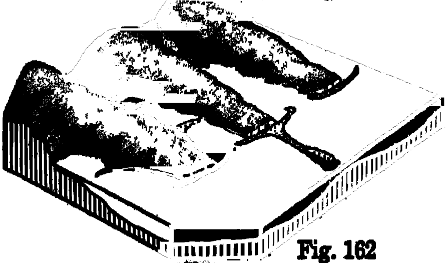


Fig. 162

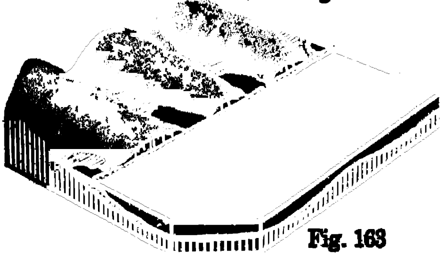


Fig. 163

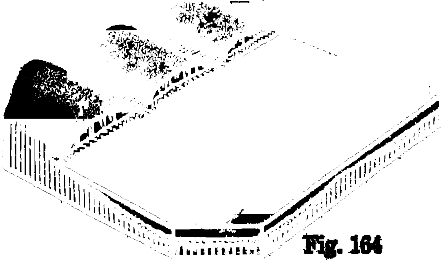


Fig. 164

FIGS. 160-164. Evolution of one type of shoreline on a bold embayed coast. (*Modified after D. W. Johnson.*)

FIG. 160. Initial condition marked by deep bays, headlands, and islands.

FIG. 161. Cliffing of headlands.

FIG. 162. Formation of beaches and spits; an island has been tied to the mainland by a bar.

FIG. 163. Formation of truncated headlands connected by bars, forming a much-simplified shoreline. The shortened bays are being filled with stream deposits and vegetation.

FIG. 164. Headlands and bays destroyed and shoreline simplified to the greatest possible extent.

terrace. As the bench and terrace widen, the force of the waves reaching the shore becomes correspondingly feebler.

Old age is attained when the sea has cut so far inland that the waves spend most of their force in friction upon the shallow bottom as they roll across it and so make but a feeble attack upon the shore. After this stage is reached, further retreat proceeds very slowly, but the wave-cut bench gradually widens.

Low-Plain Coasts. Figure 165 shows a quite different kind of coast—a low plain devoid of bays and peninsulas and having extremely shallow water offshore. Such a coast results from the spreading of great quantities of alluvium into

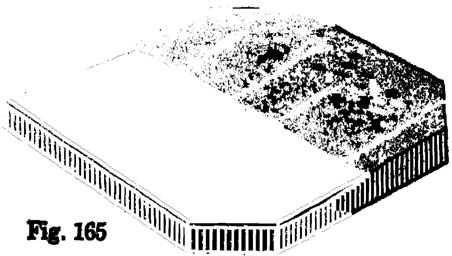


Fig. 165

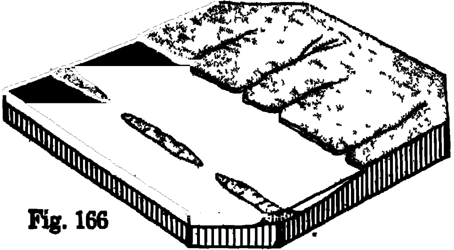


Fig. 166

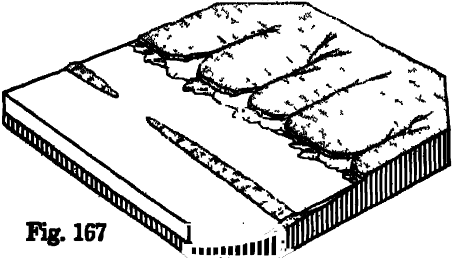


Fig. 167

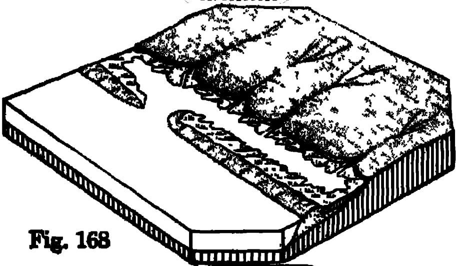


Fig. 168

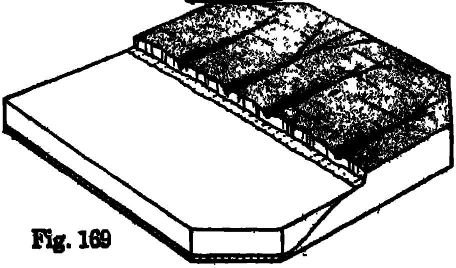


Fig. 169

FIGS. 165–169. Evolution of one type of shoreline on a low-lying plain coast. (*Modified after D. W. Johnson.*)

FIG. 165. Initial condition, marked by excessively shallow water offshore.

FIG. 166. Formation of offshore bar and lagoon.

FIG. 167. Shoreward migration of the offshore bar and beginning of filling of the lagoon.

FIG. 168. Continued migration of the offshore bar and conversion of much of the lagoon into tidal marsh.

FIG. 169. Removal of the offshore bar and tidal marsh, permitting the waves to attack the mainland directly.

the sea faster than the waves can attack and remove it, as happens where a great river like the Mississippi is building a delta. It may result also from emergence that brings a broad area of shallow sea floor out of water. As a result of the shoal water, the waves drag bottom well offshore and pick up sediment which is dropped at the line of breakers, gradually building an offshore bar (Fig. 166). Built up above sealevel by storm waves, the bar forms islands inclosing a lagoon.

The sea floor outside the bar is gradually excavated by the waves, the eroded material is added to the bar, and the profile is converted into a graded profile. The waves are then able to break with force against the bar itself, eroding its seaward face. Part of the resulting debris is carried out to sea, and during storms part is thrown over the bar into the edge of the lagoon. By this process of erosion of its seaward face and deposition on its lagoonward face, the bar is gradually shifted landward (Fig. 167). At the same time longshore currents may add to the length of the bar by building out one or both of its free ends.

Meanwhile the lagoon, rarely more than 20 feet deep at the start, is slowly filled with sediment washed into it from the land, as well as with the debris thrown over the bar. Where the water is shallow enough, vegetation thrives and adds its quota to the accumulating sediment, locally forming extensive tidal marshes. As the bar is driven landward the lagoon becomes a tidal marsh throughout its extent (Fig. 168). Eventually both bar and lagoon filling are completely cut away as the breaking waves continue their attack upon the land. The removal of the offshore bar ends the stage of youth (Fig. 169).

After the disappearance of the offshore bar the sea launches its drive against the shore with unchecked vigor. A wave-cut cliff and bench are rapidly formed, and the shoreline develops minor irregularities controlled by the distribution of weak and resistant rocks. The remainder of the process is identical with that of the stages of maturity and old age on the embayed coast already described. The most noticeable result is the production of an ever-widening wave-cut bench.

The significant features in the evolution of this type of graded coast, as compared with that of the bold coast first described, are (1) the creation of a profile, chiefly by erosion, from an initially too-shallow bottom, and (2) the absence of bay bars. It is evident that, on any type of coast, the rate at which the changes occur must depend on the resistance of the rocks and the strength of the waves. Thus, if we assume equal wave strength, weak-rock shorelines may evolve to maturity while near-by resistant-rock shorelines have progressed only to the stage of early youth.

Much of the Gulf Coast of North America is a low-plain coast of this general kind. In most places it has already progressed to the stage of extensive bar building (Figs. 157, 158, 159).

CLASSIFICATION OF SHORELINES

The two examples given above are outstanding, but they are only two of many types of coast, each of which must be analyzed in the light of its own peculiarities. However, most coasts are modifications of one or the other of the types illustrated. Among them might be mentioned straight high rocky coasts free of bays and headlands (essentially the coast shown in Figs. 61–63) and low-lying plain coasts deeply indented by shallow estuaries. The evolution of each will be determined in part by the rocks of which it is composed and by the form it has at the start, but the sequence of forms developed on each will follow the principles implied in the two examples described.

The great variety of coastal types has led to many attempts at classifying their shorelines. Two different classifications are widely used. One emphasizes shore*lines*, is based on movements of the land with respect to sealevel (and vice versa), and is fourfold:

1. *Shorelines of submergence*, on coasts that have been depressed by crustal movements, or against which the sealevel has risen.

2. *Shorelines of emergence*, on coasts that have been upheaved by crustal movements, or from which the sealevel has fallen.

3. *Compound shorelines*, which are combinations of (1) and (2) above.

4 *Neutral shorelines*, on coasts which have undergone no apparent change of position with respect to sealevel.

The other classification emphasizes the coast as a whole, is based primarily on the extent of modification of the land by shore processes, and is twofold:

1. *Primary* or *youthful coasts*, whose form is principally the result of nonmarine agencies.

2. *Secondary* or *mature coasts*, whose form is primarily the result of marine agencies.

Probably no classification can be wholly satisfactory, because none can be flexible enough to allow for the many complexities evident in the history of most coasts. Nearly all coasts have been affected by repeated upward and downward movements of the land relative to sealevel and require detailed study not only of their form but also of their superficial mantle and underlying rocks before they can be fully understood.

Effect of Changes of Level. Most existing shorelines are youthful (such as those of most of New England), some are mature (such as those of the weak-rock parts of the English channel), but few if any are at present in old age. These facts are in large part the result of pronounced crustal movements and considerable fluctuations of sealevel that have taken place within recent geologic time. In particular, the shifts of sealevel incidental to the glacial ages that have occurred within the last million years have played a conspicuous part in determining the configuration of present-day shorelines.

The stage of old age in which a large island or part of a continent is reduced by marine erosion to a broad wave-cut bench fringed by a broad wave-built terrace is therefore a matter of inference. We have no evidence that either the crust or the sealevel has ever stood still long enough to permit this stage to be developed on a large scale. Since a very slight uplift suffices to bring a great deal of land out of water, shifting the zone of wave attack and forcing the process of destruction to begin over again, we must regard the process of coastal evolution as a thing that practically fails of completion.

On the other hand a broad wave-cut bench, after emergence and erosion by streams, so closely resembles the peneplane resulting from the completed fluvial cycle (p. 104) that it is difficult to distinguish between them, and we must admit the possibility that some peneplanes are, at least in part, of marine origin.

Examples of Coasts with Complex Histories. The shoreline of the Atlantic coast of North America furnishes much evidence as to its history. Although between Florida and Cape Lookout, North Carolina, there is little indication of submergence, yet, northward from Cape Lookout to the Maritime Provinces of Canada, the stream valleys are submerged, becoming more deeply so toward the north. The seaward parts of the Roanoke, James, Potomac, Delaware, Hudson, and Connecticut valleys are submerged; even Long Island Sound appears to be a stream valley that has been submerged throughout its length. As corroborative evidence, the depth of water upon the continental shelf deepens toward the north, and the shelf itself widens in the same direction.

On the other hand a narrow strip of coast, 20 to 30 feet above sealevel, from Florida to New Jersey is covered with fine-grained deposits containing marine fossils, and wave-cut features occur also at higher altitudes. Thus we have evidence of both emergence and submergence along this coast, and from the differential character of the submergence we may conclude further that the latter was brought about at least in part by differential downwarping of the coast rather than by the uniform rise of the level of the sea.

The Pacific coast likewise yields evidence of both emergence and submergence, though evidence of emergence—many wave-cut cliffs and benches (Fig. 170) lying as high as 1500 feet above sealevel—is the more conspicuous. Some of the benches are warped, showing that local movements of the crust, rather than changes of sealevel, were at least in part responsible for bringing them to their present positions. In fact the great heights attained by shore features and marine deposits on this coast reflect the mobility of the crust along the western border

SPENCE AIR PHOTOS.

Fig 170 Coast of southern California at San Pedro Hills, showing wave-cut benches high above present sealevel, emerged as a result, at least in part, of local upwarping of the crust A particularly distinct emerged bench and cliff are shown forming Portuguese Point, in the left foreground A conspicuous cliff and bench are developing at the present sealevel

of North America, whereas the character of the Atlantic coast reflects much less disturbed conditions.

Parts of the Gulf coast are bordered, like the Atlantic coast, with slightly emerged shore features, but near the Mississippi the present shoreline is somewhat embayed, indicating submergence, possibly as a result of subsidence of this part of the coast under the weight of the vast load of sediment dumped into the Gulf here by the great river. West of the Mississippi, the Gulf coast seems to have a very nearly neutral shoreline, consisting apparently of the old deltas of the Brazos and Trinity rivers and other streams.

These examples indicate the complexity of most coasts and serve to show how the study of shore features leads to important inferences regarding the history of the lands to which they belong.

ISLANDS AND CORAL REEFS

Islands. The sea contains both islands and shoals. Many of these are the tops of hills, mountains, and larger land masses that have been submerged. Most of the islands off New England and the British Isles were so formed. Others, including Bermuda, the Azores, many of the smaller West Indian islands, and the majority of the smaller Pacific islands including the Hawaiian group, are volcanic cones built up on the sea floor. Smaller islands are sometimes formed by the rock residuals left by wave erosion, and by deposits in the form of bars.

Coral Reefs. In the warm shallow water around many tropical and subtropical islands, especially in the Pacific, coral reefs abound. They are estimated to have a combined area of half a million square miles, and the calcareous debris resulting from their erosion by waves is spread over a much greater area of the sea floor. Corals (polyps), other shell-forming animals, and lime-secreting marine plants build these elaborate reefs of calcium carbonate which they secrete as limy skeletons outside their bodies. The corals reproduce by budding in plant-like fashion, so that colonies of many thousands of individuals, each living in its own tiny protecting chamber like a cliff dweller in a city of stone, combine to build a structure of great size.

As the colonies expand and build upward, the older organisms die, and gradually the dead base is buried by the growth of new generations and by the debris broken by storm waves from the parts above. The accumulating deposit is eventually cemented into a white spongy limestone upon whose upper surface the living organisms flourish (Fig. 171).

The resulting structure is a *coral reef*. Its upward growth ceases when its surface reaches sealevel, for most of these organisms can stand exposure to the air for only a few hours at a time. Over the reef the waves swirl and break, and at its outer edges, where oxygen, calcium carbonate, and food material are most abundant, corals thrive best.

Coral reefs are sharply limited in their distribution by the depth and temperature of the water, because the reef-forming organisms thrive only where the water is clear, shallow, and warm. They prefer temperatures above 68°F., and they live where the water is less than about 200 feet deep.

Saville Kent.

FIG. 171. Growing corals, visible at low tide, on the Great Barrier Reef of Australia. View looking across the lagoon toward the mainland 20 miles away.

Types of Reefs. According to their position and form, coral reefs have been grouped into three general classes: *fringing reefs*, *barrier reefs*, and *atolls* (Figs. 172–174). Fringing reefs lie close against the shore, forming platforms exposed only at very low tide. Barrier reefs lie at some distance from the shore, separated from it by shallow lagoons. Breaks in these reefs are kept open by the tides, so that the deeper lagoons make excellent harbors. Many of the high volcanic islands of the Pacific as well as islands in the Caribbean region are girdled by such encircling reefs. The west coast of the island of New Caledonia has a barrier reef 400 miles long, and the eastern coast of

Australia has one 1200 miles long. The latter lies 20 to 30 miles from shore, and its great lagoon has a depth of 100 to 200 feet. Atolls are ring-shaped reefs inclosing circular lagoons instead of islands. Generally there are breaks in the reefs, which afford access to the lagoons.

Fringing reefs are formed by organisms attaching themselves to the rocks in the warm shallow water near the shore. Barrier reefs and atolls are generally considered to have evolved in three ways:

(1) By heightening of fringing reefs, as the islands about which they were formed subsided gradually below sea-level. A fringing reef is built around a volcanic island, which begins to sink. Since corals thrive best on the seaward margin of the reef, it is only on that margin that the corals can grow fast enough to counteract the sinking of the island and keep themselves in shallow water. Continued sinking drowns the inner part of the reef, forming a lagoon protected by the growing outer fringe and causing the original fringing reef to become a barrier reef. As subsidence and erosion continue, the volcanic island finally disappears below sealevel, and the reef becomes an atoll.

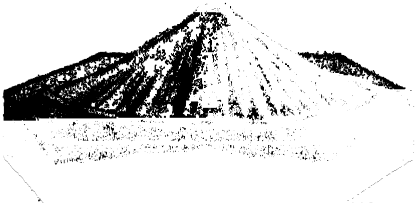


Fig. 172

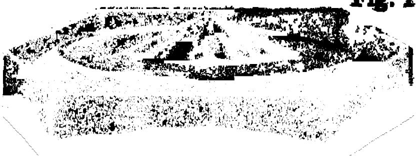


Fig. 173

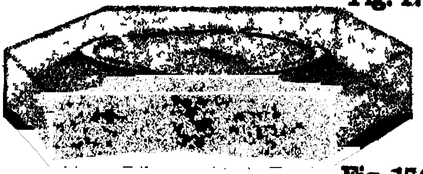


Fig. 174

Figs 172–174 Types of coral reefs.

Fig. 172. Fringing reef growing around the margin of a volcanic island.

Fig. 173. Barrier reef grown upward from a fringing reef during either sinking of the island or rise of sealevel. The reef now incloses a lagoon.

Fig. 174. Atoll grown upward from a barrier reef during further submergence of the island.

(2) By reef building on wave-cut benches during rise of sealevel. According to this view, the lowering of sealevel, caused by the removal of sea water to build glaciers during the glacial ages, destroyed preexisting reefs, checked by lowered temperatures the further growth of corals, and brought about the emergence of many wave-cut benches on tropical islands. With the later warming of the sea and rise of sealevel that accompanied the shrinkage of the glaciers, coral growth on the benches was resumed, especially on their seaward margins where

conditions are most favorable. Upgrowth kept pace with rise of the sealevel, thus forming the present barrier reefs and atolls.

(3) By upbuilding of reef structures on shallow submarine banks and island-fringing platforms not more than about 200 feet below sealevel (the normal depth range of the principal reef-forming organisms). For such growth, neither subsidence of islands and banks nor rise of sealevel is required. The reefs are slowly built up toward sealevel, where, favored chiefly at their seaward margins, they form atolls and barrier reefs.

Probably some of the existing barrier reefs and atolls have evolved in each of these ways, and some by a combination of two or all of them.

THE SEA FLOOR AND THE SEDIMENTS ON IT

Deep-Sea Basins and Continental Shelves. The enormous depressions that contain the greater part of the sea are the largest individual features of the Earth's crust (Fig. 2). Not only is their combined area more than twice that of the continental masses, but also they attain a maximum known depth of more than 6 miles and an average depth of 2.5 miles, whereas the average height of the continents (above sealevel) is only about 0.5 mile. So vast is the volume of the deep-sea basins that, if the continental masses were planed down and their debris dumped into the basins, leveling the Earth to a smooth spheroid, the sea would cover the entire Earth to a depth between 1 and 2 miles. Even this depth, however, is slight in comparison with the diameter of the Earth. If, for example, a globe 3 feet in diameter were dipped into water and then withdrawn, the film of moisture adhering to it would represent to true scale a sea one-half mile in depth. If, in drying, the globe should warp so slightly as to lessen its diameter at any place by only one-hundredth of an inch, the change would correspond to the depth of one of the deep-sea basins.

From this emerges the important fact that a very slight warping of the Earth's crust as a whole produces very great changes in the volumes of the deep-sea basins. Downwarping of their floors greatly increases their capacity and draws water away from the continental shores. Upwarping of the basin floors notably decreases their capacity and forces the water to rise and creep landward across the lower parts of the continents. The presence of widespread marine deposits high and dry upon the land, and on the other hand the presence beneath the sea of deposits that were clearly made on land, suggest that upwarping and downwarping have occurred repeatedly throughout the history of the Earth.

The term *basin* suggests a concave depression, but the deep-sea basins are depressions only by comparison with the continental masses, for because they are very shallow compared with the Earth's radius their floors are actually convex (Fig. 2). At present the sea more than fills its basins, and, spilling over them, it floods the margins of the continental masses, covering the continental shelves with shallow water (Figs. 2, 175). The shelves are widened partly by continuous erosion of the land by the sea and partly by the deposition of sediment swept seaward by marine currents. Along stable coasts where apparently these processes have operated with least interruption through a long time, the shelves are broad. Thus the shelf off the Atlantic coast of North America is 60 to 80 miles wide in the latitude of the Carolinas. The shelf off the unsteady Pacific coast, however, is in many places less than 10 miles wide, perhaps because throughout recent geologic time this part of the continental mass and the adjacent deep-sea floor have not stood still long enough to allow a broad shelf to be made.

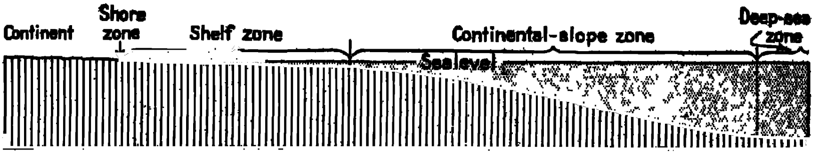


FIG. 175. The four marine depth zones at the margin of a continental (Vertical exaggeration 10 times.)

Topography of the Sea Floor. The continental shelves slope seaward at an average rate of 10 or 12 feet per mile; if the water were drawn off, they would appear to the eye as monotonous plains marked by only minor irregularities. However, the shelves (whose greatest depths range from about 350 feet to more than 600 feet) are separated from the deep-sea basins by conspicuous *continental slopes* with inclinations that average 200 to 300 feet per mile. The floors of the deep-sea basins are marked by *rises* and *deeps*. One conspicuous elevation, the Mid-Atlantic Ridge, consists of a broad ridge of almost continental proportions that trends north-south through the Atlantic Ocean basin. The deeps are not in the central parts of the basins, but lie commonly near the continental margins, some of them parallel with coastal mountain ranges. They appear to be areas that have been depressed by

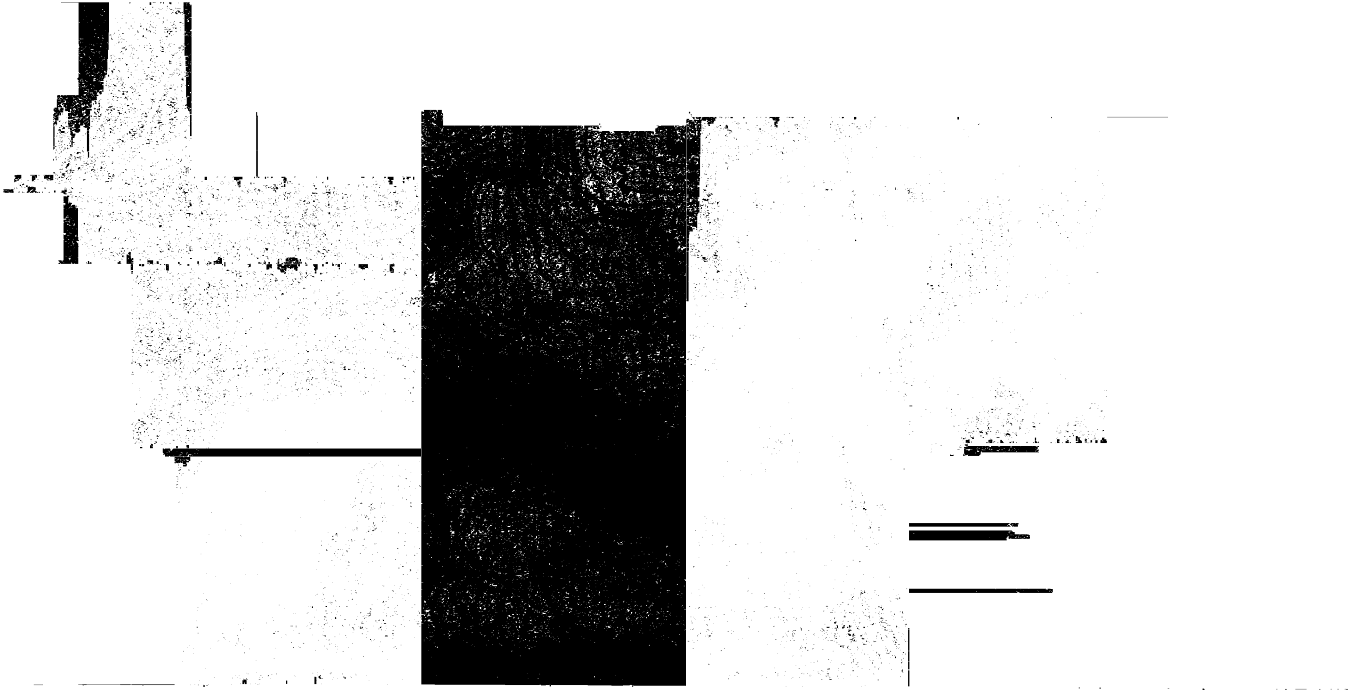


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Fig. 176. Monterey Canyon, a conspicuous submarine valley off Monterey Bay, California. The contour interval is 300 feet; the canyon opens into the Monterey Trough at a depth of 9000 feet. Other submarine valleys are visible in the continental slope.

breaking or sharp bending of the deep-sea floor and probably are related to uplifted continental margins. The Pacific Ocean holds the record for depth with 35,400 feet at a point east of the Philippine Islands. In the Atlantic, which in general is shallower than the Pacific, the Milwaukee Deep off Puerto Rico holds the record with 30,246 feet. These great depresssions in the sea floor correspond in area and in magnitude to the highest elevations on the land. The greatest deep lies more than 6.5 miles below sealevel, and the loftiest mountain chain, the Himalaya, stands 5.5 miles above it.

Formerly, these broad rises and deeps were believed to be the chief topographic features of the sea floor, which was conceived to be throughout most of its extent a monotonous plain with gentle slopes and very little local relief. This belief resulted from the fact that, until recently, soundings were made from exploring ships by the process of paying out wires to which weights were attached. This operation was so laborious that soundings were few and too widely spaced to detect much local relief. Within recent years, however, there has been developed a method whereby sound waves are transmitted from a ship to the bottom, from which they are reflected and are picked up again on the ship by an instrument that automatically calculates and records the depth. The position of the ship is determined by surveying methods, and in this way an area of sea floor can be charted quickly.

Sonic surveying of this kind has revolutionized our picture of the sea floor. Instead of the plains-like surfaces that were once believed to be nearly universal, broad areas of the floor are now known to consist of plains and high mountains, and the continental slopes are seen to have an intricacy of detail rivaling that of complex land surfaces. In some places the detail appears to have resulted from local warping and faulting of the crust (Chap. 15) and submarine volcanic activity, but in others it consists of valley systems not unlike those that diversify the land (Fig. 176). Geologists are not agreed as to whether these features are valley systems cut by land streams and later submerged, or depressions excavated in some manner by submarine currents.

Sea-Floor Sediments. No two parts of the sea floor are exactly alike in the character and distribution of their sediments, but certain general factors apply to them all. Slope, distance from shore, amount of sediment available, currents, temperature and oxygen content of the water, and depth are the principal controls. Of these controls depth may be the most important, for it determines the effectiveness of waves and currents in distributing rock waste worn from the lands, and at the same time, by controlling the sunlight and heat that penetrate to the sea floor, it regulates the habitats of rock-forming sea animals and plants.

The rock waste derived from the land is in more or less constant transit seaward, until bit by bit it comes to rest upon parts of the bottom too deep to be reached by waves and currents, and there it remains as long as those parts of the sea become no shallower. Therefore ideally grain size should decrease with increasing depth and distance from shore. While this transfer and deposition of rock waste are in progress, limestone is being built up in other places that are warm, shallow, and free from land-derived sediment (Chap. 12).

Shore Zone. The shore zone, the area between the levels of high and low tide, emerges twice every day. Its sediments vary from place to place and from time to time. They range from gravel along steep, cliffed coasts to sand on many beaches and bars, and to silt and clay in protected bays and lagoons. In exposed places the sediments are shifted about, raked over, and sorted by every change in wave strength. The general mantle of rock waste in transit seaward does not form a layer uniformly thick, for it includes conspicuous shore-zone accumulations such as beaches and bars, whereas in protected bays and lagoons, finer sediments accumulate.

Shelf Zone. The zone of the continental shelves is one of change and activity. Waves and currents keep the water in motion, salinity and turbidity vary from place to place, and the temperature changes with the seasons and varies with latitude. Hence the sediments vary greatly from place to place. Along the bold and exposed coast of California the near-shore sediments include gravel and coarse sand; off the mouth of the Mississippi the deposits are finer sand and mud derived from far in the interior; and bordering the low and nearly streamless limestone coast of southern Florida are limy deposits of very fine texture.

Discussion of the graded shore profile brought out the fact that all the rock waste handled by waves and currents is in a state of intermittent transit seaward, toward resting places too deep to be affected by the motion of the water. Hence we might reasonably expect that the average grain size of shelf-zone sediments should decrease with increasing distance from the shore. Indeed it was formerly believed that this ideal gradation commonly occurred. However, modern study has proved that ideal gradation is rare—that the distribution of coarse and fine sediments is patchy and bears little relation to present depths and distances offshore. The inference is difficult to escape that fluctuations of sealevel during the glacial and interglacial ages have played a large part in bringing this about. During the glacial ages the continental shelves all over the world emerged widely, and the time that has elapsed since the last glacial age has been insufficient to permit the waves to reshuffle the sediments into the ideal gradation seaward from the present shores.

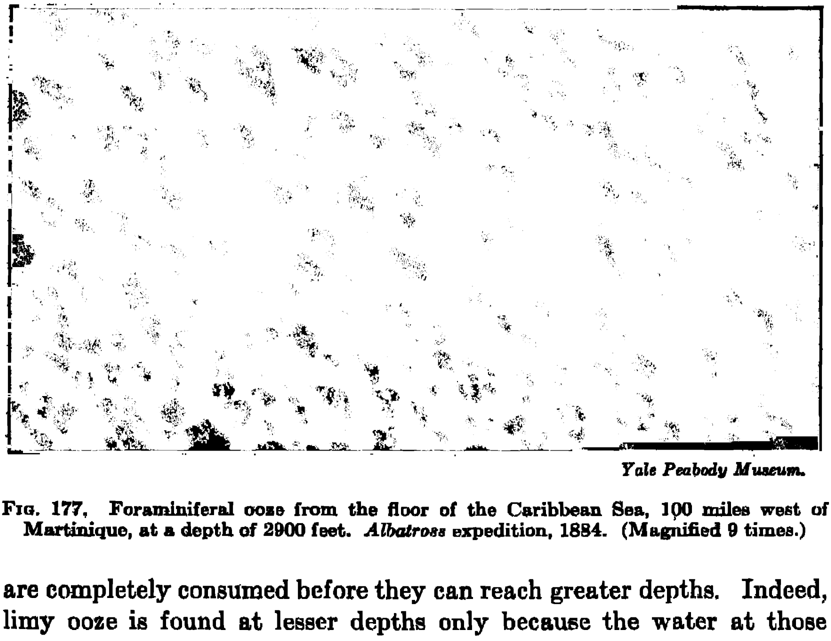
✓ *Continental-Slope Zone.* On the great slopes leading from the continental shelves down to the deep-sea floors only the finer-grained kinds of land-derived sediments are found. These include wind-blown as well as sea-borne detritus. In some areas there occur also the oozes characteristic of the deep-sea zone.

Deep-Sea Zone. Thousands of samples of sediments from the vast area of the deep-sea floor have been brought up from the depths by ingenious scoops let down on wires miles in length, from vessels specially equipped for marine exploration. Since 1935 there have been developed various types of bottom-sampling apparatus that can be lowered to any depth. These consist essentially of a metal tube, 10 feet or more in length, that can be driven into the sea floor. The tube automatically closes and is hauled to the surface, bringing with it a core sample of the uppermost few feet of sea-floor sediments.

Further use of such apparatus will add enormously to our knowledge of deep-sea sediments. The information we have at present indicates that vast areas are mantled with extraordinarily fine, slimy deposits known as *oozes*. Their thickness is unknown, but it may be great. The oozes are of several kinds and are derived from various sources. *Red clay* is the most widespread variety, occurring commonly at depths exceeding 13,000 feet. Apparently it consists of fine wind-blown dust (derived both from erosion of the land and from volcanic eruptions), pumice (p. 309) that has floated for a time before sinking, tiny fragments of meteorites that have fallen directly into the sea, and also the insoluble residues of the shells of organisms that have sunk to the deep-sea floor. Most of this material is thoroughly altered by chemical decay; in fact its red color probably results from oxidation during its accumulation at the slow rate of about 1 inch in 3000 years. Another variety is *blue mud*, a land-derived fine silt and clay that forms a belt surrounding the continental masses. Because it accumulates more rapidly than the red clay, oxidation of it is incomplete; therefore the sediment retains its blue-gray color instead of turning red.

Other large areas of the sea floor are underlain by organic oozes formed largely of minute shells and shell fragments dropped by organisms that live near the surface of the deep sea. One, the most widespread, is *foraminiferal ooze* (Fig. 177). It is formed largely of the tiny calcareous shells of Foraminifera, single-celled animals which float in myriads near the surface in the warmer parts of the sea. In the process of reproduction the parent emerges from its shell and subdivides into many daughter cells, each of which in turn secretes a new shell and repeats the life history of its parent. As reproduction is rapid, the abandoned shells drift down like a perpetual snowfall into the depths.

Foraminiferal ooze rarely occurs at depths exceeding 16,000 feet, because the limy shells, dissolving as they drift very slowly downward, are completely consumed before they can reach greater depths. Indeed, limy ooze is found at lesser depths only because the water at those depths is more nearly saturated with calcium carbonate, and hence solution takes place slowly.



Yale Peabody Museum.

FIG. 177. Foraminiferal ooze from the floor of the Caribbean Sea, 100 miles west of Martinique, at a depth of 2900 feet. *Albatross* expedition, 1884. (Magnified 9 times.)

Radiolarian ooze is formed by the Radiolaria, another group of floating microscopic animals, which, unlike the Foraminifera, fashion their delicate and ornate shells from silica. *Diatom ooze* is formed by the accumulation of the remains of diatoms, microscopic plants, which also secrete minute shells of silica.

The deep-sea floor is therefore a huge area in which the finest of all sediments are quietly and continuously accumulating. These sediments are added to by submarine volcanic eruptions. In high latitudes they are augmented also by the dumping of rock fragments rafted seaward on floating ice. Layers of such "glacial" sediments alternating with foraminiferal oozes in core samples from the North Atlantic sea floor suggest climatic fluctuations within the recent geologic past.

Permanence of the Deep-Sea Basins. The oozes brought up from the great depths give evidence that has an important bearing on the history of the Earth. The sands, muds, and limy deposits of the shelf zone are gradually converted into sedimentary rocks. These rocks are identical with many of the sandstones, siltstones, mudstones, and limestones found widespread in all the continents. For this and other reasons it is certain that these rocks of the continents were formed in shallow seas, many of them epeiric, at various times when the seas overspread wide areas of lowland. In spite of the many broad marine invasions, no continent has yielded a sedimentary rock proved to be a solidified deep-sea ooze, although such rocks do occur on certain islands. As far as this evidence may be relied on, we infer that, for as long a time as is recorded by the rocks, the continental masses have existed as features distinct from the deep-sea basins. The major relief features of the globe appear therefore to be very ancient and more or less permanent structures.

Deposits of the Shelf Zone as Sedimentary Rocks. In examining the sandstones, siltstones, shales, and limestones that constitute an important part of the rocks of the continents, we can see the record of repeated marine invasions of lowlands, many of the invasions having formed epeiric seas. The compositions of these rocks reveal the character and positions of the former coasts and the depths of water offshore; and the relationships of the shelf-zone rocks to shore-zone deposits formed at the same time enable us to locate the actual positions of former shores whose expression on the surface of the land was obliterated long ago. These matters are discussed in Chapter 12.

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A presentation of the view that barrier reefs and atolls originate chiefly by island subsidence.

(*b*) The Changing World of the Ice Age; by R. A. Daly. Chapter VII, Coral Reefs and the Ice Age. Yale University Press, 1934

A presentation of the view that barrier reefs and atolls have originated chiefly by glacially controlled fluctuations of sealevel

(*c*) The Antecedent-Platform Theory; by J. E. Hoffmeister and H. S. Ladd, Journal of Geology, Vol. 52, pp. 388–402, 1944.

A presentation of the view that barrier reefs and atolls may originate without relative change of sealevel with respect to the land.

A group of three professional papers that discuss the origin of coral reefs. They should be read as a group.

11. Recent Marine Sediments. A Symposium; edited by P. D. Trask. 736 pages. Thomas Murby & Co., London, 1939.

A comprehensive treatise by a group of authorities.

CHAPTER 12

SEDIMENTARY ROCKS

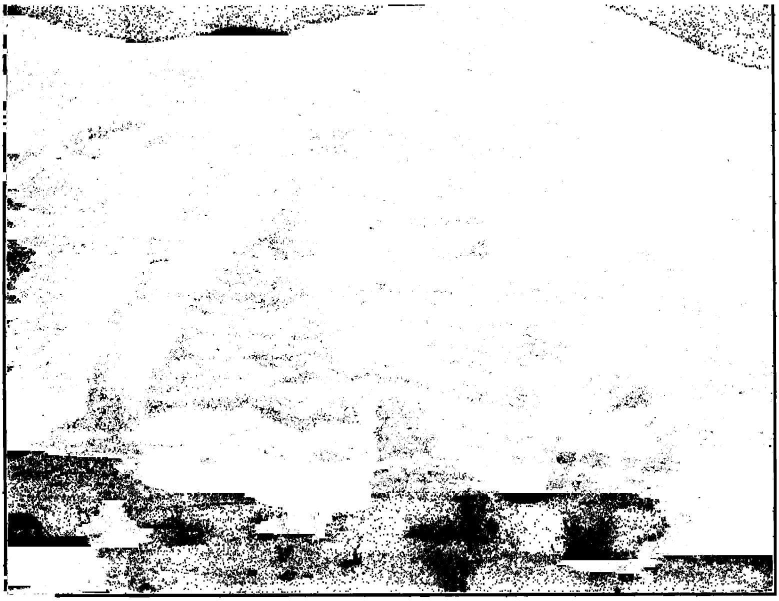
FUNDAMENTAL IMPORTANCE OF THE SEDIMENTARY ROCKS

The larger part of the Earth's land surface is underlain by sedimentary rocks; in fact, they make up four-fifths of the total land surface of the globe. Igneous and metamorphic rocks constitute the remaining fifth. Because the sedimentary rocks are so widespread, they determine to a large extent the varied aspect of our landscapes. They underlie the vast expanses of the Great Interior Plains of the United States; in the northern Rocky Mountains they are sculptured into the towering cliffs and peaks of Glacier National Park; and in the Southwest they are magnificently displayed in the Grand Canyon of the Colorado. Furthermore, they contain stored within themselves the mineral fuels coal, oil, and gas, as well as many other valuable mineral resources. This fact in itself assures their great practical interest to man. To the science of geology they are fundamental, because they comprise the most informative pages of the book in which the long history of the Earth is written. They record the extent of former seas, the advance of these seas over the continents and their subsequent retreat, the uplift of continents, and the birth of mountain ranges and their ultimate destruction. To indicate in a measure how we decipher these records is the main purpose of this chapter.

SEDIMENTATION

The processes now at work on the Earth in wearing down the rocks, in transporting the debris, and in depositing the resulting sediment are explained in previous chapters. "Sediment" ordinarily means the fine-grained material that has settled out of a liquid; but as used in geology it has a far wider meaning. Not only sediments that were laid down in bodies of water, but also materials deposited by wind and by ice are included. Much coarse material also is included, as for example glacial and alluvial deposits containing boulders many feet in diameter.

Sediments now being deposited or deposited so recently that we know the conditions under which they were laid down can be seen at many localities. In an alluvial fan, for example (p. 87), abandoned stream courses radiate from its apex and give us vertical sections into the material of which the fan was built. These sections reveal the nature of that material and the details of its arrangement and structure (Fig. 178). Recent delta deposits also can readily be examined. A reservoir may for some reason be drained; it soon becomes gullied by rainwash, and the sediments that have accumulated on its floor can then be seen in the excellent cross-sections cut by the gullies. The cores of deep-sea sediment brought up from the bottom of the North Atlantic Ocean in recent years (p. 249) give us vertical sections as much as 10 feet long of the material now accumulating there. The sediment contains shells of organisms that are living in the surface waters as well as those living on the muddy sea floor. Consequently we know the environmental conditions in which these sediments are accumulating as well as the nature of the sediments themselves. To these examples,



C. R. Longwell.

Fig. 178. Constitution and structure of an alluvial fan as exposed in the steep side of a stream valley. The largest boulders are 5 or 6 feet across. Grand Wash, Arizona.

as well as to many others, we apply the uniformitarian principle of Lyell that the present is the key to the past (p. 9). The facts obtained by examining sediments now being laid down under known conditions help us to interpret how sediments were formed in the past.

The problem considered in this chapter is the inverse of that considered in the earlier chapters, which explain the processes now visibly operating on the Earth. The problem now before us is one frequently met in geology. It is this: Given a sedimentary rock, what can we infer as to the conditions that prevailed during the deposition of the sediment from which the rock was formed? Whence came the material that makes up the rock? Was the material carried to the place of deposition by water, wind, or ice? In short, how much information can we elicit from the given sedimentary rock?

To answer this question we must determine the mineral composition of the given rock; what structures are produced by wind, water, or ice as transporting agents; what features indicate that the sediment was laid down in the sea or on the continents; and how the sediment was converted into rock.

ORIGIN AND KINDS OF SEDIMENTS

In preceding chapters we have seen how erosion tends to reduce the land area of the globe to sealevel. The products of this ceaseless attack are twofold: one, mechanical, comprising bits of rocks and minerals, collectively called *detritus* (dĕ-trī'tŭs); and the other, chemical, comprising dissolved substances such as calcium carbonate. These are carried away by a transporting agent, and eventually the detritus is dropped by streams, wind, or ice as a mechanically deposited sediment. Dissolved matter is carried to the sea or to interior bodies of water, such as Great Salt Lake and the Dead Sea, where part of it is eventually precipitated in consequence of some chemical or physical change in the body of water. Conceivably therefore there are three classes of sediments: (1) detrital[1] sediments that have been deposited mechanically from the transporting agent; (2) chemical sediments formed of compounds that either have been precipitated from the solvent or else have been withdrawn from the solvent by the physiologic processes of plants and animals; and (3) sediments formed by combination of the two modes of deposition.

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[1] Detrital, fragmental, and clastic are synonymous.

MECHANICALLY DEPOSITED SEDIMENTS

Detrital Sediments. Detrital sediments are classified chiefly on the basis of the size of their dominant particles as *gravel, sand, silt,* and *mud* or *clay.* Gravel is a coarse sediment comprising boulders, cobbles, and pebbles. Sand is composed of grains smaller than gravel, generally like granulated sugar in size. Silt and clay are composed of the very finest-grained products of erosion, so fine-grained in fact that, unlike sand grains, they cohere when wet.

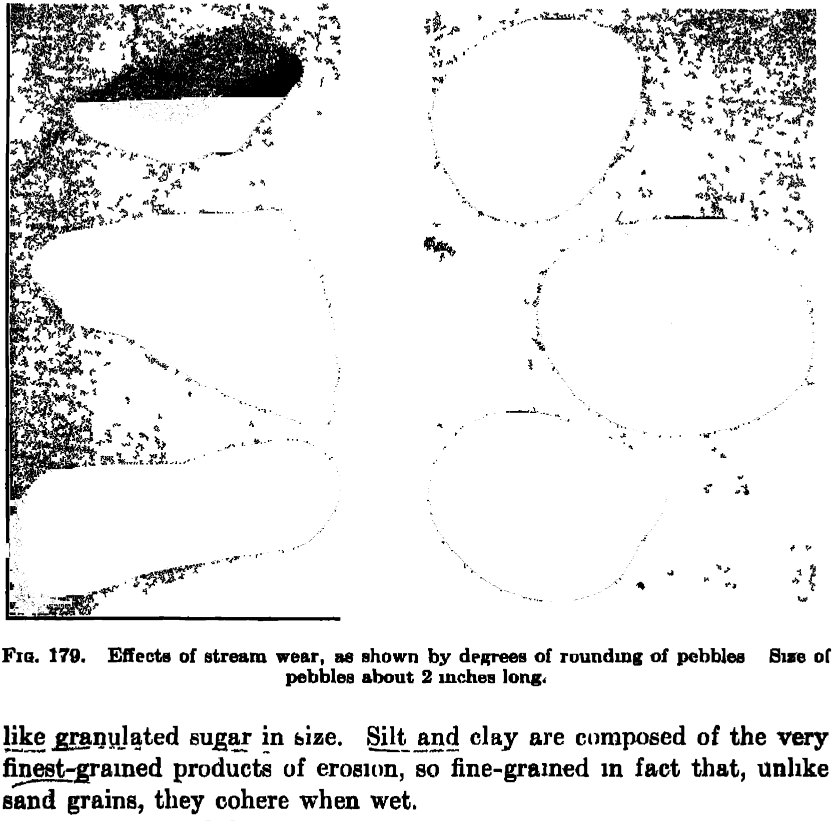


FIG. 179. Effects of stream wear, as shown by degrees of rounding of pebbles. Size of pebbles about 2 inches long.

Sand grains and the coarser detritus in gravels are more or less round. At its source the detritus consists of irregular, angular pieces of rock bounded by joints or fracture surfaces, but as the result of impact and abrasion during transport the fragments lose their edges and corners. The farther they travel the more rounded they become (Fig 179). Perfectly homogeneous rock fragments become spheroidal or spherical. Fragments having planes of weakness, such as cleavage or foliation, become ovoids or flat disks. Therefore angular and subangular fragments in a gravel indicate that they have been carried no great distance from their parent rock. As the pebbles of softer and less coherent rocks wear away faster than the more resistant fragments, durable materials, such as quartz and rocks composed of quartz, and tough coherent rocks, such as basalts, predominate in well-rounded gravel. On the other hand, gravel consisting of pebbles that have not been transported far may contain less durable minerals and rocks, such as feldspar, schist, and limestone. Limestone fragments rarely occur in gravels because, being soft, they are readily abraded and, being soluble, they tend to be dissolved. In the gravels of arid regions, however, limestone fragments are common because of the scarcity of running water to abrade and to dissolve them.

Large sand grains become more thoroughly rounded than small grains. The smaller grains, cushioned by the buffer action of the film of water surrounding each grain, become rounded with difficulty or not at all, as is shown for example by the fact that most of the grains deposited at the mouth of the Mississippi River are angular despite their long transport. They are below the size at which rounding by water is effective. In general river sands are more angular than lake or marine sands. Wind-blown sands are the most conspicuously rounded, and in the so-called millet-seed sands, common in some deserts, the grains are perfect spheres whose dull-lustered surfaces resemble ground glass, the result of natural sandblast action.

Quartz is the commonest constituent of sand because of its great hardness and chemical indestructibility. Because of the prevalence of quartz, "sand" has commonly come to mean quartz sand. "Sand" refers only to the grain size of the particles, and many sands contain rock fragments and minerals other than quartz, such as feldspar. On the beaches of coral islands the "sand" particles are made up of broken bits of coral and other organic remains.

In silt and especially in clay, quartz is present in but minor amount, and finely flaky minerals predominate. The reason for this predominance is that during transport the flaky minerals, because they cleave readily, become comminuted to extremely minute flakes, and these minute flakelets, because they tend to float, are slow in settling to the bottom. They are so minute, in fact, that most of them can not be certainly identified even with the most powerful microscope. Consequently in recent years the more potent method of identification by X-ray analysis is being used, and, latest of all, they are being studied with the electron microscope, which gives magnifications up to 100,000 times. In this way the mineral composition of even such minutely grained aggregates as clays is at last being established.

Clay has the supremely important technologic property of being moldable, or plastic, by virtue of which it can be fashioned when wet into any desired shape and retains this shape on drying. This plasticity is caused by the content of flaky minerals and by the fact that the flaky minerals composing clay are surrounded by films of water, which act as a lubricant and facilitate slipping of the flakes into new positions when a mass of clay is forced to change its shape.

Clay, as the product of the deposition of the finest detritus, has a wide range of composition. Its most characteristic components are the flaky minerals already mentioned that are mainly hydrous silicates of aluminum,[1] but include also white mica and chlorite.

PHYSICO-CHEMICALLY FORMED SEDIMENTS

There remains now to account for the fate of the matter taken into solution during erosion of the land. Every square mile of the 40 million square miles of the Earth's surface that drains to the sea contributes yearly 70 tons of dissolved matter, mainly calcium carbonate and silica, but also other substances. In part this material is withdrawn from the sea by chemical precipitation, as by the action of some reagent or by some physical change such as rise of temperature, as well as by the physiologic processes of plants and animals; in part it remains dissolved and accumulates in the sea, thus increasing the saltness of the sea slowly, exceedingly slowly in fact. Many times in the geologic past, arms of the sea have evaporated sufficiently to cause the dissolved salts to be thrown down as extensive beds of rock salt. In the Mid-Continent region salt beds estimated to contain 30,000 billion tons of salt were laid down. Vast as such deposits are, they are mere bagatelles compared with the enormous deposits of calcium carbonate that have been laid down by the sea (1) by inorganic processes, such as precipitation caused by a decrease in the carbon dioxide content of the sea water; and (2) by the organic agencies of plants and animals whose vital processes withdraw from solution the chemical compounds that are fixed in the organisms. By photosynthesis algae extract carbon dioxide from water; in consequence, the calcium bicarbonate in solution is changed to the highly insoluble compound calcium carbo-

[1] Called the "clay minerals," chiefly kaolinite, montmorillonite, illite, and halloysite.

nate, which is therefore precipitated on the plant surfaces. Bacterial activity also has been supposed to cause the precipitation of calcium carbonate from water carrying calcium bicarbonate in solution. Corals and other lime-secreting organisms extract calcium carbonate from sea water to form their skeletons or shells. The silica dissolved in the sea is kept down to a vanishingly small amount because it is being steadily extracted by silica-secreting organisms, chiefly diatoms, radiolaria, and sponges (p. 250).

PYROCLASTIC SEDIMENTS

When the material blown out during volcanic eruptions settles out of the atmosphere, it forms *pyroclastic* sediments (Chap. 13). Because of their volcanic origin they differ from ordinary *clastic* sediments, which are made of material derived from the destruction of older rocks by weathering and erosion. Pyroclastic rocks comprise many extensive deposits. The finer pyroclastic material (the so-called volcanic ash) forms the most widely continuous deposits known. A single mighty eruption, such as that of Tamboro in 1815, can spread a continuous deposit over an area of 1,000,000 square miles.

The rocks formed by the consolidation of these sediments are called tuff and volcanic breccia. They can be regarded as igneous rocks because they are formed by igneous activity; on the other hand they can be regarded as sedimentary rocks because they are made of fragments that accumulated according to the laws of sedimentation. Here they are treated as sedimentary rocks, but they form an exception to the rule that sedimentary rocks consist of materials derived from the erosional destruction of older rocks.

SEDIMENTS OF SPECIAL TYPES

Wind-blown sand and silt; glacial detritus, such as till; and carbonaceous materials, such as peat, are sediments of special types. They are of much interest and importance, both practical and scientific, but in bulk they fall far behind those already mentioned.

KINDS OF SEDIMENTARY ROCKS

Naturally the origin and kinds of sedimentary rocks are determined chiefly by the nature of the sediments from which they are formed, for a sedimentary rock is merely a sediment that has been converted into rock. The principal detrital sediments and resulting sedimentary rocks are, according to a somewhat simplified scheme, as follows:

SEDIMENTS	CONSOLIDATED EQUIVALENTS
Gravel	Conglomerate and breccia
Sand	Arenite Sandstone, made up of quartz fragments Calcarenite, made up of calcareous fragments
Silt	Siltstone
Mud	Lutite Argillaceous (shale) Calcareous (calcilutite)
Pyroclastic sediments	Tuff and breccia

The most practical classification of sedimentary rocks at present is based on chemical composition. On this basis they fall into the following classes:

I. Siliceous rocks.
II. Argillaceous rocks.
III. Carbonate rocks, mainly limestone and dolomite.
IV. Other rocks, e.g., phosphate rock, iron-ore rocks.

CONVERSION OF SEDIMENTS INTO ROCKS

Sediments as initially deposited are generally loose, incoherent, and highly porous. They become converted into rocks by compaction, by the deposition of cement in the pore spaces, and by physical and chemical changes in the constituents. Not all these processes necessarily affect every sediment. As the result of any or all of them the sediment is made firmer, harder, more coherent—it has become *consolidated* into a rock.

Compaction reduces the initial large porosity of clay sediments. A growing mass of sediment increases in thickness, and the upper beds press more and more heavily on those under them. In clays this sedimentary loading crowds the grains together, diminishes the pore space, and squeezes the water out of the pore space. A freshly deposited clay has a porosity of 50 per cent or more. By the time the clay has been buried 1000 feet it has become compacted to 80 per cent of its initial volume; and at 5000 feet it has been reduced to one-half its initial volume. Such highly compacted clays have become shales; that is, they no longer become plastic when mixed with water. Although compaction greatly increases the density of clay, probably a chemical change involving dehydration of the constituent clay minerals is the actual cause that changes the clay into shale. The sedimentary loading is accompanied by a rise of temperature, the result of the geothermal gradient, and although the rise is small it probably suffices to dehydrate clay minerals and to destroy their plasticity.

Another process potent in converting some sediments into rocks is due to the property of gelatinous constituents—the mineral jellies (the hydrates of silica, alumina, and iron)—to change spontaneously soon after coagulation, whereby they lose water and harden. Silica jelly (or "gel") when freshly precipitated is a soft, gelatinous mass; on aging, as the chemists say, it hardens into opal, and the opal may later crystallize into chalcedony. Changes of this kind, though affecting only the mineral jellies distributed in small amount through the mass of a sediment, can greatly increase its coherence and hardness.

Cementation is the deposition of mineral matter in the spaces between the grains of sediment. The most common cements are calcite, silica in the form of quartz or chalcedony, and iron oxide. These cements may have been deposited concurrently with the crystallization of any gels that were disseminated through the sediments, or they may have been brought in later either from the adjacent body of sediment or from afar, by ground water moving through the pore spaces.

The conversion of a sediment into rock as a result of compaction, and of the hardening caused by the aging of the gelatinous constituents, may go on practically contemporaneously with accumulation. However, it is generally postponed, and in some sediments it may never occur. The older sediments are generally more thoroughly consolidated than the younger. Formations of geologically recent origin consist as a rule of loose or incoherent material; the more ancient formations consist of hard rocks. A most remarkable exception to this general rule is afforded by the Paleozoic sediments of Russia and adjacent parts of Finland. In Karelia, Finland, is an ancient clay, of Cambrian age, which is commercially used in preference to the clays of Recent age. The Cambrian clay has remained unchanged since it was deposited as sediment on the floor of the Cambrian sea half a billion years ago and has kept its plasticity after this tremendous lapse of time. In Russia a Carboniferous limestone is so loose and crumbly that it resembles the chalk of western Europe, which is tens of millions of years younger, and the coal near Moscow, though older than the coals of the world's principal producing districts, has not advanced beyond the stage of lignite, the stage so typical of the immensely younger "brown coals" of Germany. That these ancient sediments have not been consolidated is doubtless due to the facts that they have never been buried under a heavy load of superincumbent strata and that they have remained undisturbed because they are in a region of great geologic tranquillity.

STRATIFICATION

As seen from the rim of the Grand Canyon of the Colorado, the most striking features of the view are the layering of the sedimentary rocks in the walls of the gorge and the marvelous and varied colors of these layers. Each layer is a *bed,* or *stratum* (plural, *strata*). Individual strata may differ greatly in thickness, ranging from those hardly more than a film in thickness to beds many scores of feet thick. An individual stratum is generally separated from the one below it and from the one above it by a surface of discontinuity. As a rule the rock can be readily separated along these surfaces. Such surfaces of discontinuity are called *bedding planes* or *stratification planes. Stratification* is the term applied to this arrangement in layers (Fig. 180), an arrangement well shown in the canyon cut by the Colorado River and in countless other river canyons of lesser depth. Stratification is generally easy to recognize, because strata may differ from their associates in color, composition, or texture, or in all three. The Coconino sandstone of the

C. R. Longwell.

FIG. 180. Stratified sandstone and shale. Near Morgantown, West Virginia.

Grand Canyon can be recognized at many miles' distance by the clear white color that sets it off from adjacent beds, and in the same way the Redwall limestone catches the eye by its bright red color.

Stratification is generally the most easily recognized structural feature of sedimentary rocks, but it is simulated in places by other structural features, such as joint planes and the cleavage of slates. Because stratification is fundamentally significant in determining the geologic structure of a region, ability to distinguish it with certainty from all apparently similar features is of prime importance.

Not all sedimentary rocks show stratification. Such absence of bedding planes in a rock suggests that it was formed in some of the less common ways in which sediments are known to have been deposited, as, for example, the deposition of till by a glacier.

Sedimentary strata forming today under standing water as on the bottom of the sea, or a lake or a gulf, are horizontal or nearly so. Similarly, streams that are depositing their load have a low gradient; consequently the resulting stratified sediments are nearly horizontal. Therefore, strata that are nearly horizontal, as in Fig. 180, are said to be undisturbed, that is, in the attitude in which they were originally deposited. In many places, however, we find strata that are not horizontal, but are inclined or even stand vertically. We are then forced to conclude that the strata have been tilted from their original position: in short, that they have been disturbed; in many places they have been profoundly disturbed. This conclusion, now a commonplace in geology, carries so many astounding implications that it required hundreds of years to become generally accepted.

Sedimentary rocks have been forming ever since the first rains fell on our planet. Since then some 500,000 feet of strata (100 miles!) have been laid down No such immense pile has been deposited in any one place, however The exceptionally deep gash in the Earth's crust cut by the Colorado River exposes a pile of horizontal sedimentary rocks only 4000 feet thick. The deepest well so far drilled has penetrated sedimentary beds to a depth of 17,823 feet[1] without, however, having gone through their full thickness; sensitive geophysical instruments used in exploring for deeply concealed mineral resources show that the stratified rocks extend in places to depths of 30,000 to 40,000 feet. It is a notable fact in geologic history that, wherever sedimentary rocks have accumulated during a continuous period of sedimentation to a thickness of 40,000 feet or thereabouts, the resulting pile was subjected

[1] As of December, 1947.

to powerful movements that acutely folded the strata and ended further deposition of sediments in that particular tract of sedimentation (p. 460). Forty thousand feet appears to be the maximum thickness of material deposited in any one basin of accumulation.

Stratification is commonly caused by changes in velocity of the transporting agent. An obvious example is afforded by the fine silt and clay that is deposited upon coarse gravels during times of diminished flow of water on an alluvial fan. Current velocity may fluctuate daily, as on the outwash plain fronting a glacier, where the transporting competence of the streams varies enormously with the time of day. Such variation is vividly impressed on anyone who has had to ford the braided network of streams flowing out from a glacier: low water in the cool of the morning when melting of the ice is at a minimum, and flood water in the heat of the afternoon, when the continuous bumping of boulders moving downstream can be heard.

More commonly changes in the velocity of streams are seasonal, because of floods during the rainy season and low water during the dry season.

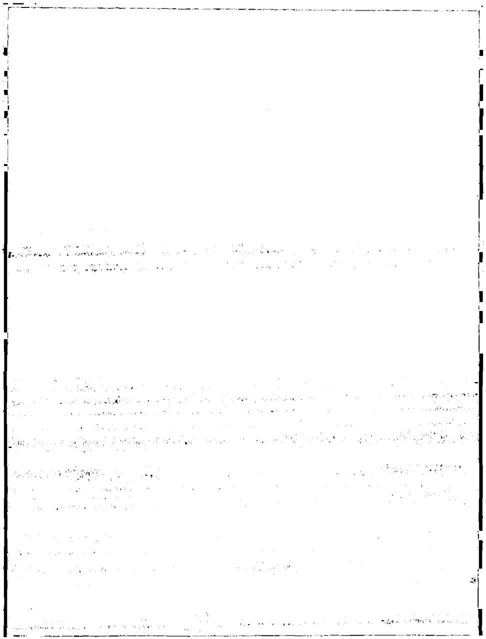
In the sea the normal arrangement of sediments is this: coarse material lies near the shore and grades seaward into fine material—muds. In times of storm, however, when a strong undertow develops, the coarse sediment is moved out into deeper water and is laid down on the muds that had accumulated earlier during periods of calm. This alternation of storm and calm produces an interstratification of coarse and fine sediments.

Another cause of stratification is a change in the kind of material that is being fed into the basin of deposition. Limestone beds alternating with shale beds indicate a marked change in composition of the material brought into the basin of deposition: the limestones record clear seas free from land-derived waste, whereas the shales record an influx of muddy sediments.

Some sedimentary rocks show a regular alternation of material of two kinds, giving rise to the so-called banded rocks, from the banding seen on their outcrops. Layering of this kind suggests the influence of some naturally occurring rhythm: to account for the thinner layers (the laminae), ebb and flow of the tide and alternations of summer and winter have been suggested. Other cycles have been suggested to account for the thicker layers (the Sun-spot cycle of 11 years, and still longer cycles), but convincing proof has not yet been found.

The summer and winter layering caused by the annual march of the seasons appears to be well developed in certain sedimentary deposits.

A deposit formed during one year is called a *varve*, regardless of whether it was formed in the sea, a lake, or elsewhere. Most varves consist of two thin laminae of differing color and composition. Such laminated sediments occur in glacial clays, in which summer, the time of high water, is represented by a silty lamina and winter by a lamina of very smooth-feeling, "fat" clay. That each couplet of silt and clay represents the deposit formed during a single year, like a tree ring, is well established, so that the couplet can confidently be called a varve. Recognizable annual deposits are now forming in some lakes; and they are thought to have been recognized in certain marine shales, in playa clays, in dune sands, and in the marine salt deposits that occur in Stassfurt, Germany, and in Texas. The regular repetition of layers made up of laminae of unlike composition is easy to recognize, but to establish that each layer represents a year is difficult, and so far conclusive evidence is not at hand for all laminated sediments suspected of being varves. The recognition of varves is a matter of great interest,



W. H. Bradley, U. S. Geological Survey.

FIG. 181. Laminated lake sediment, from the Green River formation, Colorado. Shows annual layering (varving), the light-colored laminae having been deposited during the summers and the dark laminae during the winters. (Enlarged 3 diameters.)

however, for varves give us a means of eventually establishing a chronology based on years.

A laminated lake sediment regarded as of varved origin is shown in Fig. 181; it is from rocks known as the Green River formation, in Wyoming and Colorado. Each varve consists of a pair of laminae; one lamina is composed largely of calcium carbonate, and the other is much higher in carbonaceous matter and hence much darker. During the summer the lake warmed up, consequently calcium carbonate was abundantly precipitated, and the production of floating microscopic organisms reached its annual maximum. The carbonate, being heavier than the organisms, sank promptly and formed a light-colored lamina, which thus records the summer. The organisms, composed largely of organic matter, sank more slowly, however, and so most of the organic matter settled during the winter upon the lamina deposited during the summer. The pair of laminae, thus interpreted, record a year; they constitute a varve. By counting and measuring the varves in the Green River formation, which is 2600 feet thick, it is estimated that between 5 million and 8 million years were required to deposit the formation.

COLORS OF SEDIMENTARY ROCKS

A prominent feature of many sedimentary rocks is their color. As seen in natural exposures the color generally differs from that seen on freshly fractured surfaces. The colors seen in canyon and cliff are the results of weathering and generally are only skin deep, but they enhance strikingly the picturesque qualities of such country as the Grand Canyon, the Painted Desert, and Zion National Park. As a rule, the color of the weathered rock is warmer and more brilliant than the intrinsic or proper color of the unweathered rock as shown on fresh fracture. To illustrate: the cliffs on the western slope of the Bighorn Mountains, in Wyoming, are strikingly scarlet but the rock of these cliffs when freshly broken is pure white.

The inherent color of a sedimentary rock is determined by the color of the predominant mineral of which it is composed, and by the nature, amount, and mode of distribution of the pigment in the rock. Carbonaceous matter and iron compounds are the chief pigments. They are present in small amounts, from a fraction of 1 per cent to several per cent, but the thickness of the coating of these pigments around the sedimentary grains is probably the chief factor in producing the colors of the rocks.

Carbonaceous matter colors sedimentary rocks dark gray and black. On exposure to light the carbonaceous matter bleaches, so that rocks thus pigmented are lighter in color on natural exposures than they are on freshly broken surfaces. Carbonaceous pigment originates from the alteration of organic matter that was buried with the sediment.

Iron compounds, mainly ferric oxide and ferric hydrates, containing iron in its most highly oxidized form, are powerful coloring agents. Ferric hydrates give yellowish red, brown, and maroon tones. Hematite, the anhydrous ferric iron oxide (Fe_2O_3), produces red tones. A sandstone or shale containing much less than 1 per cent of finely disseminated hematite is likely to be brick-red or vermilion. In the presence of reducing agents, such as carbonaceous matter and hydrogen sulphide, both of which are products of decaying animal and vegetal matter, ferric iron becomes reduced to the ferrous state, in which state the iron forms much less conspicuously colored compounds.

Red sedimentary rocks have long excited special interest because of their striking color. Some are associated with beds of rock salt and gypsum formed by evaporation of large bodies of salt water, and they are common as sandstones and conglomerates containing undecomposed feldspars. That the feldspars are not decomposed indicates that the rocks from which they were derived were destroyed by mechanical erosion faster than they were by chemical decay. Either a cold climate or an arid climate would favor the preservation of feldspars in unaltered condition, but so would very rapid mechanical erosion and deposition, even in the warm, moist climate of the tropics. Red beds commonly have features indicating that they might have formed in warm, arid regions, and this fact has inclined many to regard all red sedimentary rocks as having been formed under the influence of warm, arid climates. Opposed to this belief, however, is the fact that the regions in which red soil is now forming are warm and humid, whereas most deserts have dun or brownish soils. As soils are the chief sources of the sediment laid down by streams, the most prolific source of red detritus is manifestly the soils of warm, humid lands whose relief is sufficient to insure free underground movement of the oxygenated water of the rainfall.

In low, swampy tracts, vegetation is luxuriant, and the resulting abundant decaying vegetal matter causes strongly reducing conditions. Therefore, if a red soil formed in a humid region is carried by erosion into low-lying swampy tracts and deposited there, its red color will be destroyed by reducing agents, and the resulting sediment will be dark colored; consequently red deposits are not likely to form in moist regions. However, in deserts and in regions of seasonal rainfall, vegetation is sparse and the soil becomes dry during the period of drought, and consequently the humus is destroyed by oxidation. Reducing conditions are therefore absent, and if red sediment is washed in from the moister uplands it is likely to remain red.

Owing to the abundance of life in the sea, reducing conditions generally prevail on the sea floor; consequently, most marine sediments are not red. Nevertheless, where red sediment is swept into the sea in large quantity the organic matter on the sea floor may not suffice to reduce all the ferric iron. As a modern example we find that red mud is accumulating off the mouth of the Amazon and other rivers draining areas of lateritic soil (p. 48); in time these red muds will doubtless be converted into red rocks. As a matter of fact, many red sedimentary rocks contain abundant marine fossils—clear proof that these red rocks were formed in the sea.

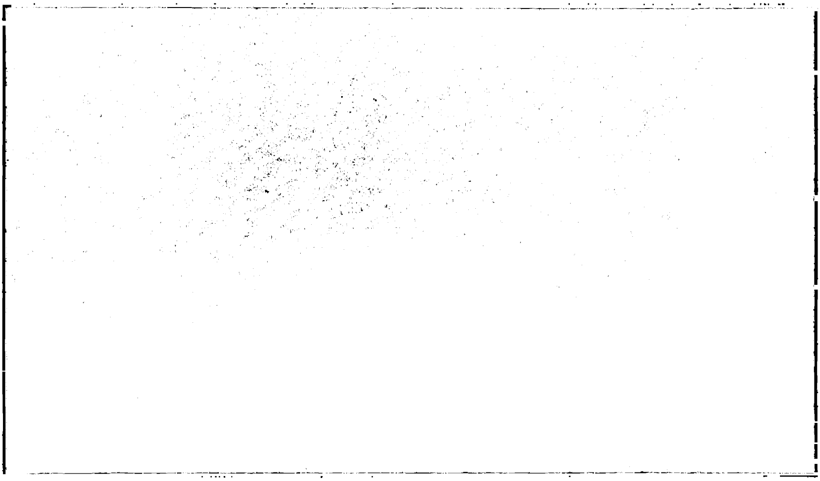
In conclusion, red strata occur chiefly in terrestrial formations, especially in those that accumulated in warm arid or semiarid regions; but redness is not in itself proof that beds so colored were deposited in any particular environment.

The problem of the origin of red beds illustrates the principle that similar rocks may be formed in several different ways. The correct interpretation of such rocks requires the gathering of all available evidence and its critical analysis. Nevertheless, in the present state of our knowledge, the evidence we can acquire may prove to be insufficient to give a definitive answer.

OTHER CHARACTERISTIC FEATURES

There are other characteristic features of sedimentary rocks not so universally prevalent as stratification and color but nevertheless of common occurrence and of great value as indicating the conditions under which the sediments were deposited.

Mud Cracks. Certain low flat places, such as a quarry floor or a shallow concrete ditch beside a highway, are favorable places for the accumulation of mud after a heavy rain. Anyone who has seen this mud after it has dried for several days in the sun and wind will notice that the mud has shrunk and cracked into highly irregular polygonal blocks. An example of unusually regular cracking is shown in Fig. 182. Shrinkage cracks thus formed are called *mud cracks*. On further exposure the blocks of mud bake and harden. Later, if there is a prolonged dry spell, wind-blown sand or silt may cover the surface and fill the cracks with sediment that is coarser than the material of the mud-cracked layer. The form of the polygonal blocks is thus preserved, even if the mud flat is again inundated. Mud-cracked layers formed in this way can later become hardened into beds of shale and sandstone, and after geologic ages become exposed to view by the work of erosion. The fossil "mud cracks" will then be revealed on the bedding planes. If the beds are in a tract that became involved in one of the revolutions that have affected the Earth's crust and have been turned so that their topsides are now down, the mud cracks give us the means to determine this fact, for in their original position the cracks tapered downward and they now taper upward.

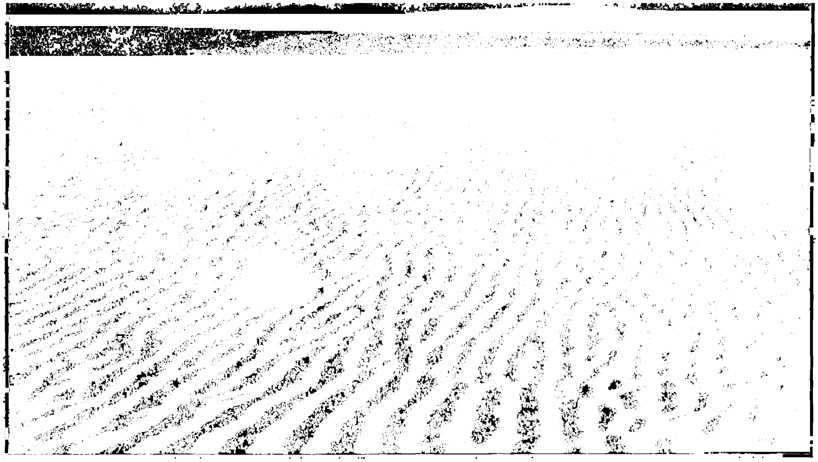


C. R. Longwell.

FIG. 182. Mud cracks formed in the playa of Pintwater valley, Nevada.

Clearly the conditions that favor the formation and preservation of mud cracks are ideal also for preserving the footprints of animals that may have walked or run over the soft mud soon after it was deposited. Footprints and mud cracks therefore commonly occur together in sedimentary rocks, although not many mud-cracked rocks show footprints. Similar conditions favor the preservation of raindrop imprints, markings that record the fall of a brief shower while the mud flat was exposed to the atmosphere. Singular reflections cross the mind of a geologist who surveys the impressions of raindrops in an ancient shale recording a transient shower that fell millions of years ago!

The most favorable places for mud cracks to form are the floodplains of large rivers, the landward portion of large deltas, and the wide flat shores of shallow lakes that shrink or disappear during the dry season. Marine tidal flats, in spite of their alternate wetting and drying, are unfavorable for forming mud cracks, because the mud has not sufficient time to dry thoroughly before the tide returns. Mud cracks form to a minor extent, however, at the upper margins of estuaries where the spring tides reach for only a few days in each month.



E. M. Kindle, Geological Survey of Canada.

FIG. 183. Current-formed ripple marks exposed at low tide, near Windsor, Nova Scotia. The current flowed from left to right.

Mud cracks can form under both subaerial and subaqueous conditions.[1] Some marine formations therefore contain mud-cracked beds. Notable examples of mud cracks formed under subaqueous conditions have been found in the rocks of Glacier National Park.

Ripple Marks. Where a current moves sediment along the bottom of a body of water the surface of the resulting deposit develops parallel ridges that resemble the ripples on the surface of a pool of water. These ridges are known as *ripple marks.* Every retreating tide leaves ripple marks along the beach. They are formed also by wind action on sand dunes. Ripple marks made by currents retain their form and migrate slowly in the direction in which the current is flowing, because sand grains are swept by the current up the currentward side of the ripple ridge and roll down its leeward slope. Current-formed ripple marks are therefore asymmetric, the currentward side having a gentle slope and the lee side a steeper slope. The direction of flow of the current is thus autographed in the form of the ripple mark (Fig. 183).

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[1] Mud cracks have been experimentally formed under water by the spontaneous dehydration of sediments of colloidal grain size, a phenomenon known to chemists as *syneresis.*

Oscillatory waves that touch bottom cause ripple marks to form by moving the sedimentary particles to and fro. Well-developed oscillation ripple marks are symmetric, thus contrasting with current-formed ripple marks (Fig. 184).

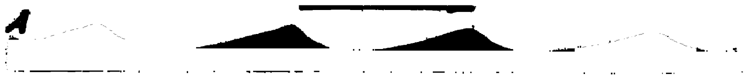


FIG. 184. Asymmetric profiles of current-formed ripple marks (*A*) contrasted with the symmetric profiles of oscillation ripple marks (*B*).

Current ripple marks are formed wherever currents flow over sandy or silty surfaces, whether the currents are winds on the land, currents in the streams, or currents in the sea. Oscillation ripple marks, however, form only under standing water, at depths not too deep to be touched by wave action. Storm waves in the sea ordinarily are ineffective below 200 to 300 feet, but exceptionally they produce ripple marking at depths of 600 feet or more.

Cross-Lamination. Some sedimentary deposits show a remarkable oblique layering, or lamination, extending diagonally across the individual beds. This arrangement in laminae inclined to the general plane of stratification is called *cross-lamination.* It is known also as cross-bedding and false-bedding. The real bedding is at right angles to the direction in which the deposit grew upward. Cross-lamination is developed whenever a sediment-laden current of air or water passes over the front edge of an embankment and spills the sediment over the slope. Anyone who watches a truck dumping dirt over the front of a fill can see how cross-lamination of this kind is formed. In nature the growing embankment may be the front of a sand dune deposited by the wind; and, in water-laid deposits, it may be the edge of a delta, the front of a gravel bar, or the front of a current ripple. The sedi-

C R LONGWELL

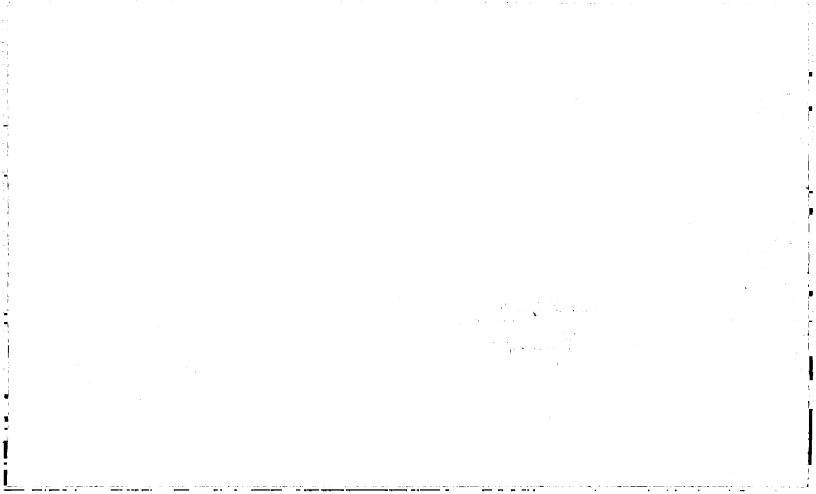
Fig. 185. Diagonal cross-lamination in sandstone. Near Zion National Park, Utah.

ment thus spilled over the slope comes to rest at an angle inclined to the general surface of deposition (Fig. 185).

The pattern of the structure produced by cross-lamination ranges from simple, in which all the cross laminae slope in one direction, to highly complex, in which the laminae slope in many directions, the result of successive truncations of the lower sets of laminae by the upper (Fig. 186). These two kinds represent the two fundamental types of cross-lamination: (1) that produced by uninterrupted deposition over an embankment, and (2) that produced by deposition interrupted by episodes of partial erosion.

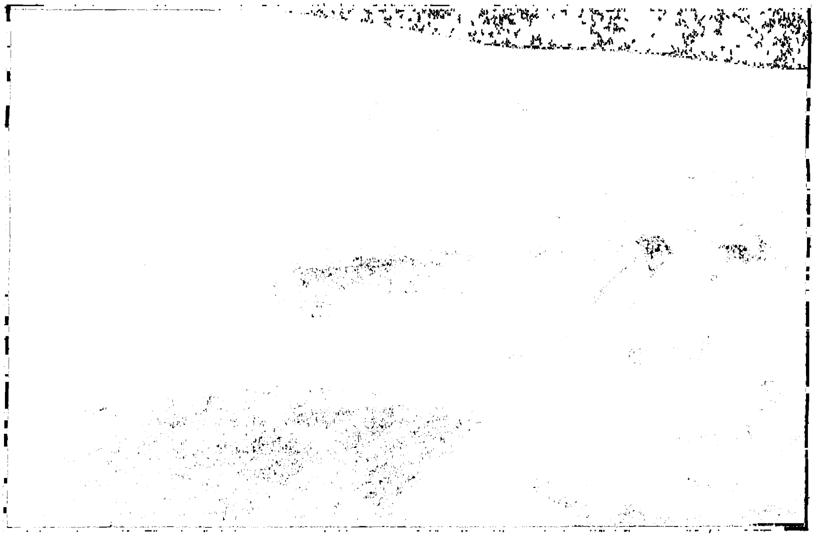
Beds that are cross-laminated range from thin to very thick. In wind-blown deposits an individual bed may be as much as 10 feet or more in thickness, but in water-laid silts the individual cross-laminated bed is generally only a fraction of an inch thick.

The cross-lamination in many dune sands is characteristic of their origin. The oblique layers are inclined in many directions within short distances, owing to the gustiness of the winds that formed the dunes. The crescentic front of traveling dunes also causes the oblique layers to slope in different directions, but a statistical evaluation of



H. E. Gregory, U. S. Geological Survey.

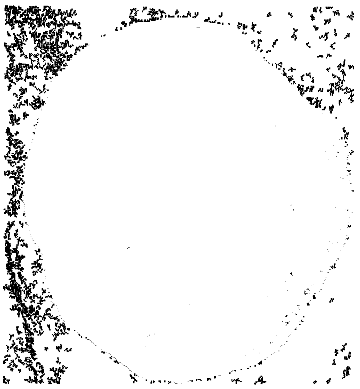
FIG. 186. Complex cross-lamination in sandstone, called festoon cross-lamination because of prevalent curved surfaces. Glen Canyon, Utah.



C. E. Siebenthal, U. S. Geological Survey.

FIG. 187. Concretions that have weathered out of the shale inclosing them. Near Havre, Montana.

many measurements will allow us to determine what was the direction of the prevailing wind. Two features useful in distinguishing eolian cross-lamination from fluvial and marine cross-lamination are: (1) few or none of the surfaces that truncate the oblique laminae in dunes are horizontal, whereas in water-laid sediments these surfaces as a rule are horizontal or nearly horizontal; (2) although most laminae are concave upward, as they are in water-laid sediments, in dunes the layers locally are convex upward. The cross-laminae that are thus convex upward represent the temporary summits of dunes formed during the upward growth of the sand deposit as a whole.



G. S. Hume, Geological Survey of Canada.

FIG. 188. Half of a ball-like concretion, which has been split in two, revealing an inclosed lobster claw.

As the cross-lamination in deposits not of eolian origin is so commonly concave upward, this feature can be used to determine the original tops and bottoms of beds that have been highly tilted or overturned by crustal disturbance.

Concretions. Some sedimentary rocks contain nodular bodies called *concretions*. These nodules differ in composition from the rock enclosing them. They are remarkably diversified in form: some are spherical or ellipsoidal, others are flat and ring-shaped, and still others are of extraordinarily odd and fantastic shapes. They range from a fraction of an inch to many feet in diameter. They are composed of one of the minor constituents of the inclosing rock; thus in chalk and limestone they consist of minutely crystalline silica and form the well-known flint nodules; in sandstone they consist of iron oxide or calcium carbonate; and in shale, of calcium carbonate or iron sulphide. The bedding planes of the inclosing strata persist through some concretions, a feature proving that these particular concretions grew in place *after* the inclosing strata had formed. Some large concretions, which have weathered out of the inclosing shale, are illustrated in Fig. 187 and show clearly that the stratification passes through them.

Manganese nodules partly encrusting sharks' teeth have been dredged from the depths of the sea, showing that some concretions form while the sediment is accumulating. But most concretions have grown in the inclosing sediments after the sediments were deposited. Many concretions contain at their center a fossil, which served as a nucleus that caused the mineral matter of which the concretion is composed to aggregate around it. Some concretions inclose remarkable imprints of fern leaves, insects, and marine animals. Nine out of every ten concretions in the Bearpaw shale of Alberta, when split in two, reveal a lobster claw (Fig. 188).

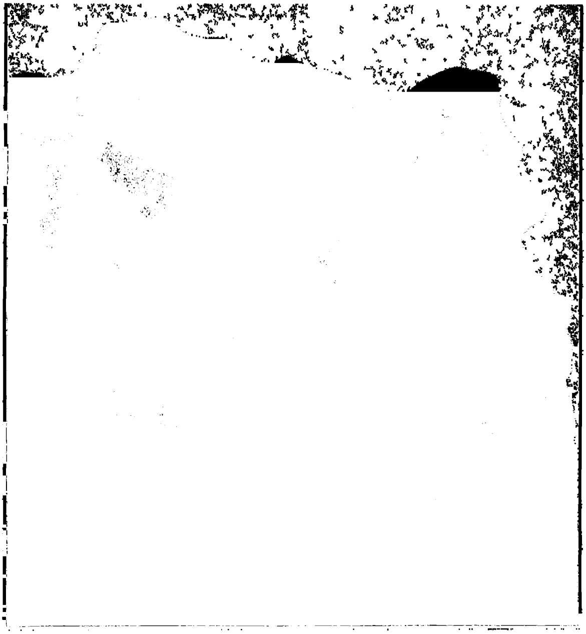


FIG. 189. Rock containing fossils of marine shells.

Fossils. Fossils (Latin *fossiles*, from *fodere*, to dig up) are the remains or imprints of animals or plants that were buried in the accumulating sediments (Fig. 189). As an adjective "fossil" has a much wider denotation, however, as shown by such phrases as fossil raindrop imprints, fossil deserts, and many more. The sedimentary rocks themselves are often characterized as "fossil sediments."

As fossils represent the animals and plants that were living while the inclosing sediments were accumulating, and generally in the same place in which the sediments were being deposited, they inform us as to the conditions that prevailed during that time. They tell us, for instance, whether the beds were laid down upon the land or under water, salt, brackish, or fresh. Fossils are in fact the chief means for distinguishing marine from nonmarine sediments. Marine sediments contain marine shells, which are readily distinguished by their characteristic features from fresh-water shells. Land-laid sediments may contain the remains of land plants and bones of land animals. Inferences as to environmental conditions during deposition, based on content of fossils, must, however, be made with discretion, because land plants and animals sometimes drift out to sea and become inclosed in marine sediments.

No organisms are preserved entire in the rocks; only the hard parts, such as shells and bones, as a rule remain. These parts may be preserved unchanged, but commonly the material of which they were built by the living organism has been dissolved and carried away by circulating solutions, and new material has been substituted so perfectly that the original form and the details of its structure are exactly preserved. Petrified wood, whose woody tissue was removed and for which silica was concurrently substituted, is a well-known example of such replacement (p. 130). Under certain circumstances the shell or hard part may even be wholly removed by water moving through the strata, and only an imprint or a hollow mold is left in the rock. A *natural mold* of this sort may later become filled with mineral matter, and this filling is a copy, or *natural cast*, of the former living organism.

Something of the climate of the region at the time of deposition can be inferred from the fossils found in the rocks. For example, fossils of palms and alligators are embedded together in certain formations in the Badlands of South Dakota, and they indicate that a tropical climate once prevailed there. Fossils are indispensable in deciphering the long history of the Earth and its inhabitants; in fact they are the fundamental basis of geologic chronology. As such they are considered in detail in books on historical geology.

STRATIGRAPHIC RELATIONS

Relative Age of Strata. The stratified rocks are formed by the deposition of layer upon layer of sediment; consequently each stratum is younger than those below it. The youngest stratum is at the top and the oldest is at the bottom of any pile of strata. Many conclusions on the structure of a region and the succession of geological events rest upon this self-evident principle.

Grouping of Strata into Sedimentary Formations. An assemblage of rock masses that have been grouped into a unit for convenience in description or mapping is termed a *formation*. A succession of strata that were deposited more or less continuously under essentially uniform conditions constitutes a sedimentary formation. A formation is usually given a geographical name based on the locality where it was first recognized, called thereafter the type locality. If the formation consists chiefly of rock of one kind, that fact is commonly expressed in the given name, such as *Bighorn dolomite, Utica shale*, or *Coconino sandstone*. If, however, a formation is made up of rocks of several kinds (for example, limestone and shale), it is given a geographical name only, as *Kansas City formation*.

A formation generally contains fossils that are distinctive, which are therefore termed *index* or *guide fossils*. By means of its index fossils a formation can be recognized wherever it occurs and can be traced from the type locality across country. The areal extent of a formation can therefore be determined and can be plotted on a map. This procedure is called geologic mapping. It is the first step in any geologic study of an area, whether the purpose is purely scientific or is essentially practical.

Formations laid down in geologically recent time contain the remains of plants and animals of species still living. Formations of somewhat older geologic age contain, along with the remains of living species, the remains of species of organisms that are now extinct. In still older formations the remains are entirely those of extinct species of organisms. A species that has become extinct never reappears in strata of younger age. By piecing together the evidence from the order of the superposed strata with the kinds of fossil organisms occurring in them, a chronology of the history of the Earth has been worked out. The larger subdivisions of this chronology are given in Appendix D.

The order in time when successive fossil faunas and floras lived and appeared in the strata and when they became extinct was first established in Europe and America. This order can now be used to determine the age and succession of strata anywhere the world over. Probably the most astounding triumphs in deciphering the geologic record by means of the fossils contained in the rocks have been achieved in regions where crustal revolutions have caused the strata to stand vertically or have overturned them, or have thrust great thicknesses of older strata over younger strata, as in Glacier National Park (Fig. 307).

Areal Extent of Sedimentary Formations. Sedimentary formations differ greatly in areal extent, depending on the conditions under which they were formed. Shales and limestones make up formations that are as much as thousands of square miles in area. For example, the Madison limestone of Montana and adjacent States has been traced over an area of more than 140,000 square miles.

Beds of clastic origin, such as sandstone and shale, indicate that there were land surfaces from which the sediments were derived by erosion and basins in which they were deposited. If the sediments were

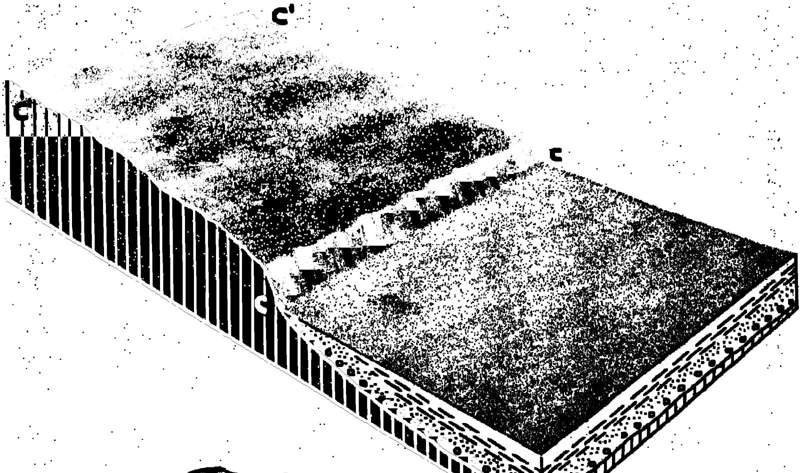


Fig. 190

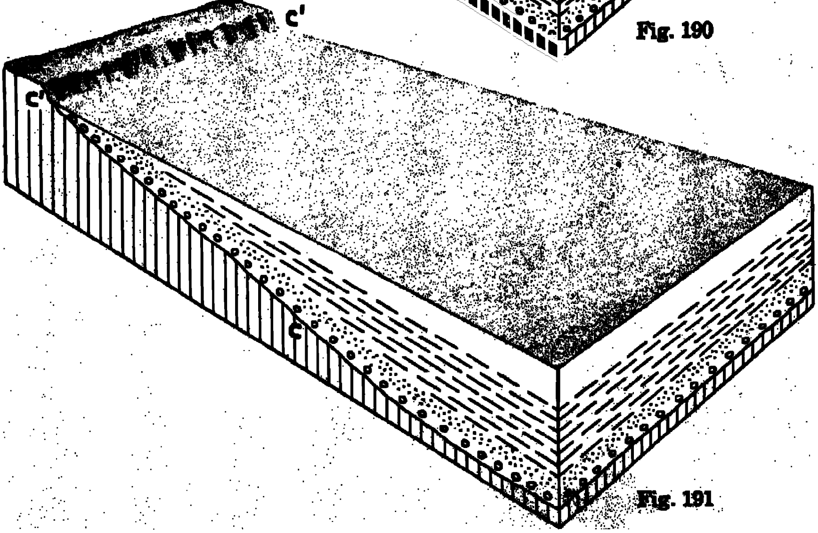


Fig. 191

Fig. 190. An early stage in the advance of the sea over the land mass shown on the left. CC represents the position of the sea cliff.

Fig. 191. A later stage in the advance of the sea, showing the land mass largely consumed. C'C' represents the new position of the sea cliff. Each of the layers of mud (which eventually become shale), represented by the broken-line pattern, extends much farther to the left than the layer below it; these layers *overlap* landward.

laid down along an open coast, the deposits tend to be wedge-like, thickest near shore and progressively thinner seaward. The sediments poured into the Gulf of Mexico by the Mississippi River are derived from a drainage area of more than 1 million square miles, but the area in which they are deposited is only about one-tenth of that size.

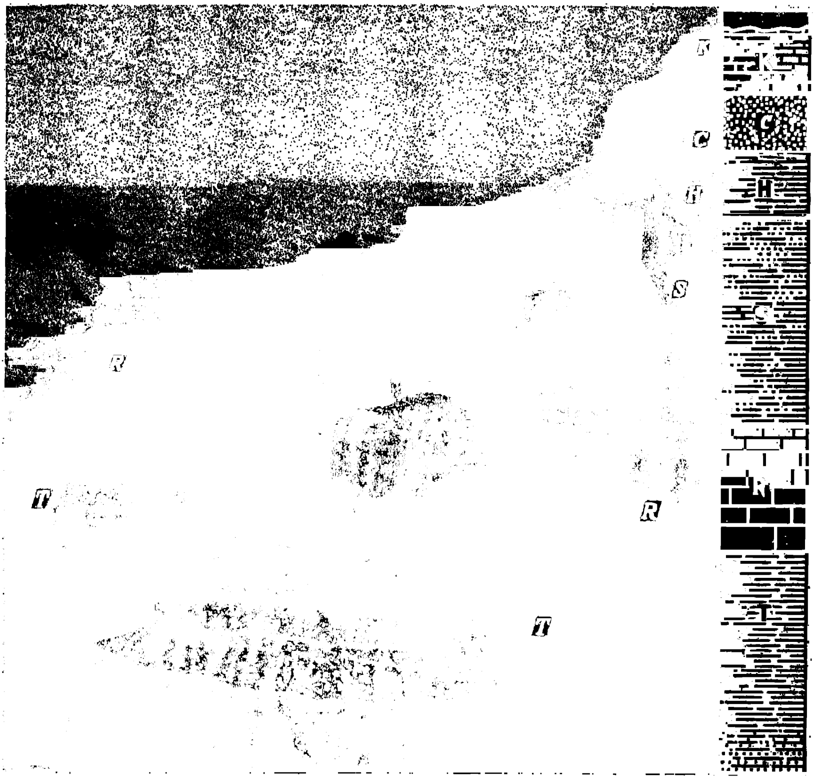
Unconformity. An important stratigraphic relation is that of *unconformity,* so important in fact that it is treated in detail in Chapter 15. A marine unconformity records that a land mass became submerged beneath the sea and a series of marine strata was deposited on it.

Overlap. When the sea advances on a subsiding land mass, the beds that are laid down on the floor of the advancing sea as the result of its attack on the land mass have the relations shown in Figs. 190 and 191. Each successively higher bed extends farther in the direction in which the sea advanced over the land mass than the bed immediately below it. Each younger bed thus laps over beyond the limit to which the underlying bed extends. This stratigraphic relation is therefore termed *overlap.*

Gravel, sand, and finer material are formed by the attack of the advancing sea on the subsiding land mass. The gravel and sand are deposited near shore, but the finer material, of mud-grade size, is carried out into deeper water. Consequently, the gravel and sand grade seaward into mud. Because the shoreline moves landward as the sea advances, what was earlier the locus of deposition of coarse material becomes the locus of deposition of fine material; therefore the sand and gravel grade upward as well as seaward into mud. Gravel of the kind shown in Figs. 190 and 191, after it has been changed into rock, is known as a *progressive marine conglomerate;* it records the advance of the sea over a land mass. As it occurs at the bottom, or base, of a formation, it is called also a *basal conglomerate.* Manifestly, the conglomerate at C' is younger than it is at C by the length of time it took the shore line to migrate from C to C'.

GRAND CANYON OF THE COLORADO AS AN ILLUSTRATION OF STRATIGRAPHIC PRINCIPLES

The Earth gives us a deep insight into her sedimentary cover in the Grand Canyon of the Colorado, where the river has cut a mighty gorge through 4000 feet of horizontal strata. Consequently the view from the rim of the Grand Canyon affords an unrivaled panorama of sedimentary rocks. At the very first glimpse of the Canyon the varied and vivid coloration, the horizontal layering, and the immense thickness of these rocks make an overpowering impression upon the observer. These rocks are here grouped into the following six formations. The lowest formation shown in Fig. 192 is made up of strata of shale and sandstone more than 1000 feet thick; above it is the Redwall limestone, whose brilliantly colored and massive beds stand out in 500-foot cliffs, forming some of the most striking scenery of the Canyon; the third is a thick series of sandstones and shales; the fourth consists entirely of shale; the fifth is the Coconino sandstone, a prominent white sandstone formed in an ancient desert; and the uppermost formation is the Kaibab limestone, so named because it forms the floor of the Kaibab Plateau in which the Canyon is incised.



N. W. Carkhuff, U. S. Geological Survey.

Fig. 192. South wall of the Grand Canyon of the Colorado River near El Tovar, showing six sedimentary formations. The formations are lettered to correspond with those shown in the diagrammatic geologic column on the right of the figure.

Marine fossils occur in almost all these formations, thus showing that the larger part of the immense pile was laid down in the sea. The beds have remained in their normal horizontal position and in undisturbed sequence, but their present position high above sealevel shows that despite their horizontal attitude they have nevertheless been lifted thousands of feet. The fossils in them tell us that the lowermost formation was laid down early in Paleozoic time and the uppermost in late Paleozoic time. Translated into years, this means a span of 300 million years (Appendix D). From the fossils we learn further that the formations that we see now represent but a fraction of this vast span of time, because the succession of life recorded in these beds is not a continuous sequence. Above the lowermost 1000 feet of shale and sandstone there is a gap in the life sequence, showing that for a long period before the deposition of the Redwall limestone no deposits were being laid down here, or if they were laid down they were removed by erosion before the Redwall limestone came into existence (p. 386). These intervals of time, although not represented by strata at the Grand Canyon, are represented elsewhere by great thicknesses of strata.

The horizontal strata exposed at the Grand Canyon of the Colorado do not tell the whole story of the geologic history of this region, but constitute, as it were, Book III of the geologic record. They rest with profound unconformity on a foundation built of two vastly older series of sedimentary, igneous, and metamorphic rocks. These rocks constitute Books I and II, the earlier history of the region. An account of this history is given in Part II, Historical Geology.

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CHAPTER 13

IGNEOUS ROCKS

Igneous rocks, as implied by their name, were formed at high temperatures; they are defined as those rocks *made by the solidification of molten matter that originated within the Earth.* Such molten matter, as is well known, is erupted as lava from many volcanoes the world over, and in these eruptions we can see igneous rocks actually in the making. Immensely spectacular though volcanic eruptions are, the progress of geologic science has shown that they are but minor manifestations at the Earth's surface in comparison with the vastly greater effects produced by active molten rock matter within the crust.

Molten matter as it rises from depth is more or less highly charged with gases. As it nears the Earth's surface in a volcanic vent, the pressure decreases and the gases begin to be liberated and to escape. Most of the gases have escaped by the time the liquid has solidified. The molten rock matter plus its content of dissolved gases is called *magma*. There are many kinds of magma—at least forty—differing widely in composition. To cite only two examples, from active volcanoes, the magma erupted by Vesuvius differs greatly in composition from that erupted by Etna.

MODE OF OCCURRENCE OF IGNEOUS ROCK MASSES

Intrusive and Extrusive Rocks. Igneous rocks occur as component parts of the Earth's crust in two ways: as *intrusive* and *extrusive* masses. In forming an intrusive mass, magma rose from deep in the Earth to higher levels in the crust, but it stopped in its ascent before it reached the surface. Consequently it cooled and solidified under a cover, or jacket, consisting of the rocks that make up the higher portions of the crust. The resulting body of igneous rock is an intrusive mass. Such a mass becomes accessible to our view only after it has been uncovered by erosion.

If magma reaches the Earth's surface and is discharged from an opening, it flows out on the surface, where it cools rapidly and solidifies. The body of igneous rock thus formed is an extrusive mass.

Obviously, an intrusive mass is formed under conditions that differ greatly from those under which an extrusive mass is formed. In forming an intrusive body the magma cools under a thick jacket of rocks; hence its dissolved gases can not escape easily, and because rocks are exceedingly poor conductors of heat it loses heat slowly and therefore solidifies slowly. In an extrusive body, on the other hand, the magma is drastically chilled by exposure to the atmosphere and therefore solidifies rapidly. Because of these differences in the way they were formed, intrusive rocks differ greatly in appearance from extrusive rocks.

Although intrusive and extrusive rocks are in general markedly distinct, in places they grade into each other. For example, magma that issued at the Earth's surface as a lava flow came up through a passageway from below; and, if the passageway remained filled, the magma in it eventually solidified into rock. Consequently the extrusive body—the lava flow at the Earth's surface—is connected with an intrusive mass below. The connection of an extrusive mass with its downward extension, or root, can rarely be seen, however, because the extrusive mass as a rule either covers its root or has been separated from its root by erosion, whereby the former continuity has been destroyed.

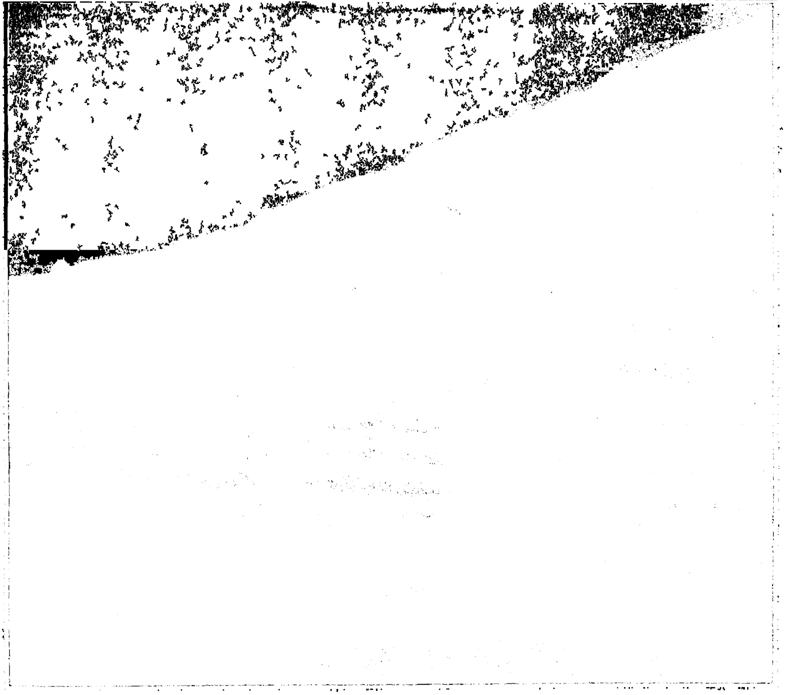
INTRUSIVE BODIES

Intrusive igneous bodies are structural units in the architecture of the crust. According to their shape, or form, and their structural relations to the inclosing rocks there are six principal kinds: *dikes, sills, laccoliths, volcanic necks, stocks,* and *batholiths.* The simplest is the dike, and it therefore will be considered first.

Dikes. A *dike* is a tabular mass of igneous rock that fills a fissure in pre-existent rocks. Consequently, a dike is inclosed between walls that are parallel or nearly so. Dikes are inclosed in rocks of all kinds—igneous, sedimentary, and metamorphic. They are said to "cut" the rocks that inclose them. By convention the term dike is restricted to those tabular masses that cut through layered rocks at an angle to the layering. If the tabular igneous mass is parallel with the layers it is called a *sill.*

Dikes range in length from a few yards to many miles; they range in thickness from a fraction of an inch to thousands of feet. Three dikes that cut a series of stratified rocks at right angles to the stratification are shown in Fig. 193.

Some fissures when formed extended up to the Earth's surface. Magma, rising from the depths along such channel ways, wells out as streams of lava. A historic example is the Laki fissure in Iceland, from which in 1783 were discharged 3 cubic miles of lava through an opening 5 miles long. After discharge at the surface ceases, the magma in the fissure solidifies; thus is formed a dike that reaches to the Earth's surface. Most dikes, however, fill fissures that did not extend to the surface, and their present exposure to our view is the work of erosion. Some dikes are the fillings of channels through which were fed other intrusive bodies, such as the sills and laccoliths next to be described.



N. H. Darton, U. S. Geological Survey.

FIG. 193. Dikes cutting horizontal sedimentary beds at right angles. Alamillo Creek, New Mexico.

During erosion, dikes that resist destruction more strongly than the rocks inclosing eventually stand out as walls; but dikes whose resistance is weaker than that of the inclosing rocks form ditches. The rock of some dikes is divided into blocks or columns by joints. These columns are perpendicular to the walls of the dike, causing the dike to resemble a pile of cordwood, an arrangement whose origin is described later under *columnar structure* (p. 372).

Dikes are likely to contain fragments of rock that were ripped off from their wall rocks and carried upward. They supply us with samples of the rocks that are present in depth and thus give us clues as to what occurs down there. Some dikes contain blocks of this kind that they have carried up from depths as great as 20,000 feet.

In many places dikes are arranged in more or less well-defined systems. Parallel dikes are in some localities so numerous and closely spaced as to constitute a dike swarm. Around certain centers, notably volcanoes, dikes have a radial arrangement, extending out from the center like the spokes of a wheel. Impressive examples of radiating dikes surround the Spanish Peaks, Colorado, and the Crazy Mountains, Montana; these mountains, however, are not of volcanic origin.

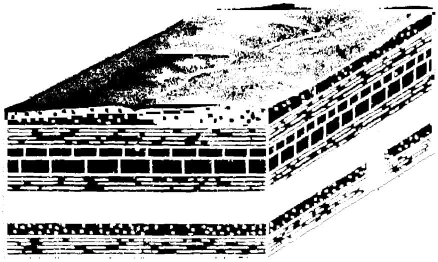


FIG. 194. The sill *SS* is intrusive between horizontal strata, the dike *D* represents the conduit through which the magma was forced up. Vertical joints were formed as the sill cooled. The block is about 1 mile wide, the sill 500 feet thick.

Dikes are adjoined by zones of rock baked and hardened by the high temperature produced by the molten magma. These effects are particularly pronounced adjacent to dikes cutting sedimentary rocks, which are more readily affected than other rocks. All intrusive masses have produced similar effects on the rocks into which they were injected. These effects give us therefore a means for determining whether an igneous mass is of intrusive origin.

Sills. A *sill* is a tabular mass of intrusive igneous rock lying parallel to the layers of the inclosing rocks (Fig. 194). Sills range in thickness from a foot or less to several thousand feet. In forming a sill, magma insinuated itself between two layers and lifted the overlying rocks through the distance now represented by the thickness of the sill. Manifestly, in order to form a thick sill deep within the crust and several thousand square miles in extent, the magma must have been injected by an enormous impelling force.

The term sill as ordinarily used means that the body so named is horizontal, like a door sill. In geology it was applied in the first place to intrusive sheets that are horizontal, such as the famous Whin sill of the north of England, which is traceable for 80 miles and averages 90 feet in thickness. The Whin sill is a mass that has remained horizontal since it was formed, but many sills have been tilted from their original horizontal position, so that by an evolution of ideas the term sill is now applied not only to horizontal bodies but also to inclined bodies. Some sills were injected into strata or other layered rocks that were already inclined at the time the magma was forced in between them. The essence of the definition of a sill, therefore, is not that a sill is horizontal, but that it is an injected mass lying parallel with the inclosing layers.

One of the best-known sills in America is the Palisade sill, which forms a long vertical cliff fronting New York City along the west side of the Hudson River. Originally the sill was 100 miles long and 1000 feet thick, but it has been somewhat diminished in size by the erosion that brought it to view. The palisade structure to which this sill owes its name is the result of the nearly vertical columns developed in it by shrinkage cracking during cooling, in the same way that columnar jointing is formed in dikes. Such columnar structures, as explained in Chapter 14, are at right angles to the walls of dikes and sills.

Laccoliths. A *laccolith* is typically a lens-shaped mass of igneous rock intrusive into layered rocks. It has a flat floor and is more or less circular in ground plan. If the magma, after it has insinuated itself between the layers in the form of a sill, arches up the overlying layers, instead of continuing to spread laterally away from the supply channel, a lens of liquid rock is produced, and this lens on solidifying forms a laccolith. The cross-section in Fig. 195 shows an ideal laccolith and the dome-shaped mountain produced by lifting and arching the rocks that overlie the laccolith.

Many laccoliths, however, depart from the typical form. Instead of being circular in ground plan, some are oval or quite irregular; and in section some have the shape of a wedge instead of the symmetrical form shown in Fig. 195. If the amount of up-arching of the overlying layers was small, the resulting laccolith approximates closely to a sill in form.

Laccoliths, more or less exposed by erosion, are conspicuous features in many parts of western North America (Fig. 196). They were first discovered in the Henry Mountains of southern Utah, where they rise above a plateau of undisturbed horizontal strata. They form a complete series that comprises all the stages of progressive denudation:

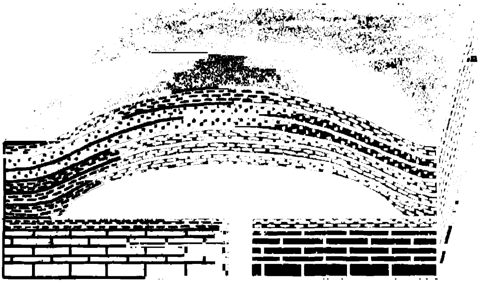
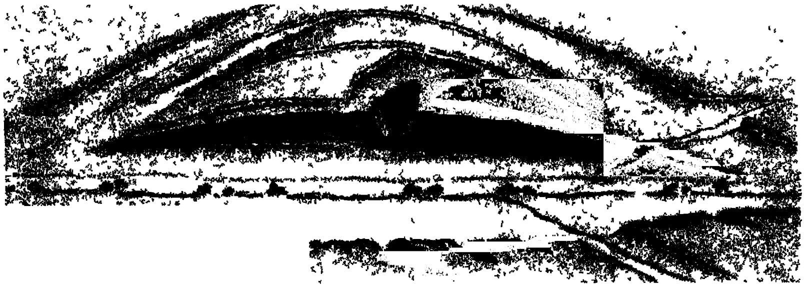


FIG. 195. Laccolith, showing the typical form of the igneous mass and the doming of the overlying beds. The block is about 2 miles wide.

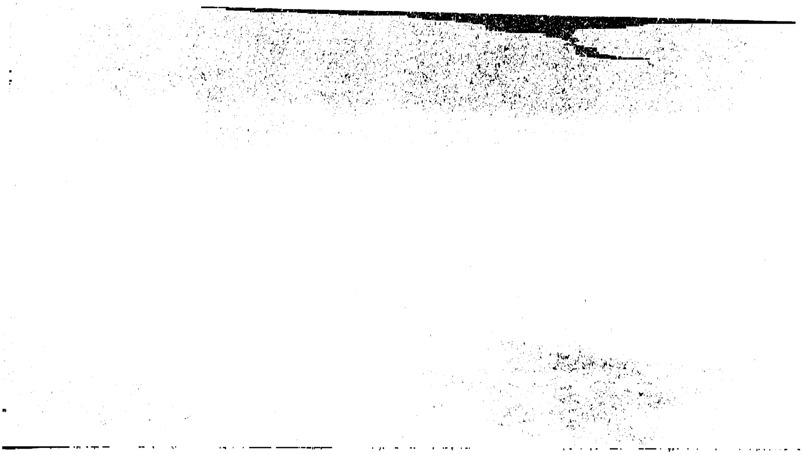
from laccoliths still covered by their roof rocks and whose presence in depth is inferred from the dome-like hills formed by the up-arching of the strata, to laccoliths so deeply dissected as to show the undisturbed horizontal strata underneath them.



Modified from L. V. Pirsson.

FIG. 196. Ideal sketch of a laccolith (*L*) from which the roof rocks have been partly stripped off by erosion. The creek cutting across the laccolith has exposed the horizontal floor on which the laccolith rests. The laccolith is about 1 mile in diameter.

Subsequently, laccoliths have been found in many parts of the world and are therefore a not uncommon form of intrusive body. Some of these later-found laccoliths are immensely larger than the classic laccoliths of the Henry Mountains, the largest of which is less than 4 miles



BARNUM BROWN.

Fig. 197. Shiprock volcanic neck, 1300 feet high, with dikes extending out from it, Near Farmington, New Mexico.

in diameter. Igneous masses as large as 6000 square miles in area have been interpreted as laccoliths.

In order that an intrusive mass be classified as a laccolith, it must have a floor. In the nature of things, the floor is the least accessible part of the mass, and, unless erosion has supplied favorable exposures, the laccolithic character of many bodies must perforce remain in doubt. Laccoliths are of great interest in theoretical geology, for they show clearly how the magma made room for itself within the Earth's crust, namely, by lifting the overlying rocks. The mechanics of intrusion of laccoliths is therefore fairly well understood.

Volcanic Necks. When a volcano becomes extinct, the supply pipe connecting it with the depths becomes filled with a solidified mass of magma or with a chaotic mass of angular blocks, large and small, termed *vent agglomerate.* This filling forms a cylindrical mass known as a *volcanic neck.* It is roughly circular in ground plan and may be from a few hundred yards up to a mile in diameter. The rocks surrounding a volcanic neck are generally cut by a radiating system of dikes, and if the rocks are stratified they are commonly injected with sills. Being in general more resistant to erosion, a volcanic neck in the course of time forms a giant tower (Fig. 197). Further description of volcanic necks is postponed to Chapter 14.

Batholiths. A *batholith* is a huge intrusive body of igneous rock. Large size and intrusive relation to its inclosing rocks are its distinguishing features (Fig. 198). Furthermore, a batholith enlarges downward and appears to extend indefinitely downward into the Earth's crust; it is without a known floor, in contrast to a laccolith, which rests on a floor.

Batholiths, like other intrusive bodies, become accessible to human observation only as the result of their exposure by erosion. Accordingly, the size of a batholith as we customarily speak of it depends on the depth to which erosion has uncovered it. The largest batholith in the United States is the Idaho batholith in central Idaho, exposed over an area of 16,000 square miles. Had erosion stripped off less of the roof rocks that formerly covered the batholith, its exposed area would be smaller; and had erosion extended a mile or two deeper its visible portion would have been much larger.

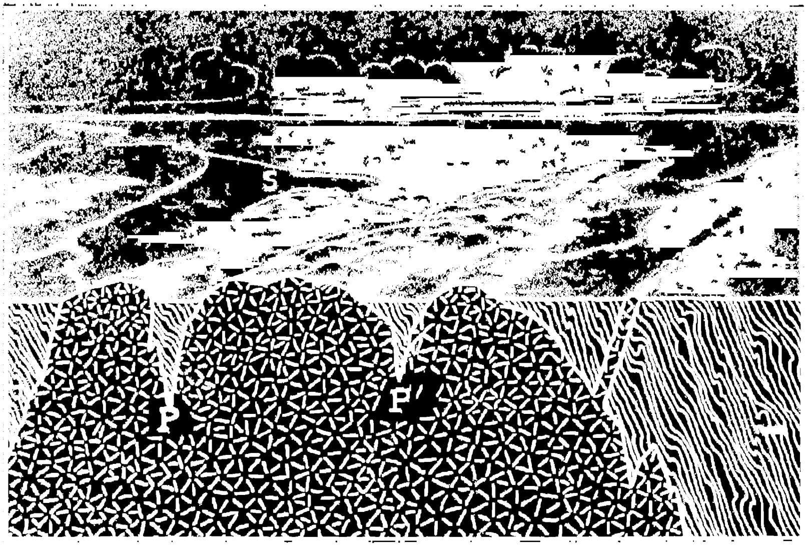


FIG. 198. Batholith, partly uncovered by erosion. *P* and *P′* are masses of invaded country rock (here shown as slate), projecting downward deeply into the batholith; they are termed *roof pendants*. At *P* the continuity with the main body of slate has persisted, but the slate mass as seen at the surface at *P′* is an "island" surrounded by granite, its connection having been sundered as the result of erosion. *S* is the top of a stock connecting with the batholith in depth. The batholith is of the *discordant* or *transgressive* type, as it cuts across the structure of the invaded country rock, the slate. The block is about 20 miles wide.

Another large batholith is that of the Sierra Nevada. It has been extensively bared by erosion, especially in the culminating portion of the range in the region of Mount Whitney, the highest point in the

United States proper. Suppose we were standing on the summit of Mount Whitney: as far as the eye could reach north, west, and south we would see granite, granite, and more granite, the solid record of what was once thousands of cubic miles of molten rock matter. Looking eastward, we would see the great granite escarpment of the Sierra dropping abruptly to Owens Valley almost below us, and we thus would perceive that the batholith is at least 8000 feet thick. Another great batholith is the Coast Range batholith, which extends for 1250 miles from the 49th parallel northward into Alaska and is 80 to 120 miles wide; it is marvelously displayed in the fiords of British Columbia and southeastern Alaska. It is about 100,000 square miles in area, the greatest batholith known, forming as it were a vast cicatrice in the crust of the Earth.

Many mountain ranges have batholithic cores, backbones of igneous rock. The home of the batholith is in the world's mountain belts, in the zones of acute deformation of the crust. The enormous bodies of magma were emplaced either at times of crustal deformation or shortly after. A batholith emplaced during a crustal revolution tends to have concordant contacts, that is, the roof rocks arch over the top of the batholith just as the roof rocks arch over a laccolith; whereas a batholith emplaced at the end of a period of crustal deformation has discordant contacts—it breaks across the general structure of the rocks it has invaded. Structurally, then, there are two types of batholith: *concordant* and *discordant*.

Granite is the chief rock of which discordant batholiths are composed, and granite gneiss (a modification of granite due to the roughly parallel arrangement of its component minerals, chiefly the biotite flakes) makes up the concordant batholiths.

As granite is formed at some depth in the crust, it is exposed at the surface only after erosion has stripped off the covering rocks; hence it is seen chiefly in those parts of the continents bared by erosion—in mountains or in regions so deeply eroded that the roots of the mountains are visible.

Granite is the main constituent of the foundation of the continental masses. These foundation granites occur as batholiths of very ancient origin, of Pre-Cambrian age, and constitute a floor upon which the sedimentary rocks of later age were deposited. Granite of younger age also occurs as colossal batholiths that are intrusive into the younger rocks, for epochs of granitic intrusion have occurred time and time again during the long span of the Earth's history. The Sierra Nevada, Idaho, and Coast Range batholiths already cited are grand examples of such younger intrusions.

From many points of view batholiths are of extraordinary interest. They are structural units of the first magnitude in the architecture of the crust. The source of the heat, the origin of the magma, the forces that set the magma in movement, and the manner in which such stupendous masses of molten rock matter make room for themselves in the higher levels of the Earth's crust are among the most fundamental and fascinating problems of geology. The concordant batholiths appear to make room for themselves by lifting and shouldering aside the rocks of the crust, like laccoliths of immense size; but the mode of intrusion of the discordant batholiths is much less clear. Many explanations have been suggested: foremost among these is the hypothesis of piecemeal stoping, so called from a certain analogy between the process by which the magma is conceived to make room for itself in the crust and the miner's process of extracting ore. The hot magma in coming into contact with the rocks it is invading shatters these rocks above the junction zone and spalls off blocks; and as soon as these blocks become engulfed in the magma they sink, because the fragments of solid rock are heavier than the liquid magma. By the progressive spalling off of these blocks and their sinking in the magma, the magmatic chamber becomes enlarged, and the magma works its way upward into the higher levels of the crust. Although it is abundantly clear that this process of piecemeal stoping has been effective around the borders of discordant batholiths, it has not yet been demonstrated that the immense chambers occupied by the batholiths were formed entirely in this way.

✓**Stocks.** *Stocks* are the smaller intrusive igneous bodies without known floors. A stock differs from a batholith only in its much smaller size; arbitrarily, an intrusive igneous body less than 40 square miles (100 square kilometers) in areal extent is called a stock; if larger than 40 square miles, a batholith. A large stock is a small batholith. Some stocks, as indicated at *S* in Fig. 198, are merely dome-like protuberances from the body of an underlying batholith.

A stock as a rule is circular or elliptical in ground plan. The distinction from a volcanic neck is not based on size, though necks are much smaller than the average stock. The term neck is used only when the evidence shows that the igneous body occupies a channel that served as the supply pipe of a volcano. Some stocks doubtless were necks, but this can not now be proved, as the evidence has been removed by erosion.

EXTRUSIVE BODIES

Extrusive bodies are those formed by magma that flows out on the Earth's surface. Magma is extruded in two ways, depending on the quantity and activity of the gases contained in it: the *quiet,* in which it is discharged as a liquid and solidifies into a massive rock, and the *explosive,* in which it is violently blown into the air and falls in the form of clots or solid fragments.

Lava Flows. Magma that issues at the surface is called *lava.* The liquid discharge, as well as its solidified equivalent, is commonly spoken of as a *lava flow.* Liquid flows are poured out from volcanoes. Some volcanoes, like those in Hawaii, discharge molten lava almost exclusively, but most volcanoes discharge alternately lava and fragmental material.

Some lava flows, however, were not erupted from volcanoes but were discharged from fissures. Such extrusion, when unconnected with volcanoes, is termed a *mass eruption.* In the geologic past such mass eruptions have occurred many times on a huge scale. Among the most notable of these enormous lava floods are those of the Columbia River region in the northwestern United States and of the Deccan region in western India. The resulting piles of lava flows in these regions are thousands of feet thick, and each covers an area of more than 200,000 square miles (p. 334).

Tuff and Breccia. According to its phase of activity, a volcano blows into the air clots of hot liquid magma or fragments of cold solidified lava torn from its throat and crater walls. The coarser pieces fall near the vent; the finer pieces, hindered by the frictional resistance of the atmosphere or carried by the wind, tend to fall somewhat later than the coarser fragments and at greater distances from the vent. On account of the highly variable intensity of volcanic explosions during an eruption, fine material is blown out during mild activity and settles near the vent, to be succeeded during violent activity by coarse material Therefore, an interlayering of coarse and fine material is produced even near the vent.

The coarser material, when converted into rock, is called *volcanic breccia,* and the finer material is called *tuff.* Both are termed *pyroclastic rocks* ("broken by fire") to distinguish them from the clastic sedimentary rocks.

Tuffs and breccias occur the world over, wherever volcanism is active or has been active. Their presence, indeed, is clear proof that volcanism was formerly active in places where it has long been extinct.

DETERMINATION OF THE AGE OF AN IGNEOUS ROCK

The geologic date at which a mass of igneous rock was emplaced in the crust or was erupted on the Earth's surface is determined from its structural relation to the adjoining rocks. Thus, an igneous body, such as a dike cutting through another igneous mass, is obviously younger than the rock it cuts. If it cuts through sedimentary strata, it is thus younger than they are. As the geologic age of sedimentary beds can generally be determined by the fossils inclosed in them, the dike is younger than the age indicated by the fossils; *but how much younger must be ascertained from other evidence.* Suppose that we have the favorable geologic exposure shown in Fig. 199. The dike is there seen to cut limestone beds, which contain fossils determinable as of Carboniferous age (Appendix D). These beds and the dike in them were worn down by erosion to a level surface, and sometime later a series of beds containing Triassic fossils was deposited on it. From the evidence thus available the conclusion clearly follows that the dike is younger than the Carboniferous beds and older than the Triassic. In geologic parlance we say that the dike is post-Carboniferous and pre-Triassic in age. Had erosion stripped off the covering beds (2), all that could be positively determined as to the geologic age of the dike is that it is post-Carboniferous.

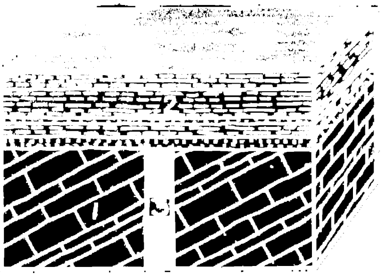


FIG. 199. Determination of the age of an igneous rock. The dike (3) is younger than the limestones (1) it intersects, and older than the strata (2) that rest on it.

Lava flows are of course younger than the rocks on which they lie. If a pile of lava flows contains intercalated layers of tuff and breccia, the layers may inclose fossils, which are the remains of plants and animals that were overwhelmed by the material blown out during the volcanic explosions. By means of these fossils the flows can as a rule be dated as to their geologic time of origin.

A sheet of igneous rock lying parallel to the strata above and below it is either a sill or a lava flow. If it has baked the strata both above and below it (p. 419), this evidence of baking proves positively that the sheet was injected at some time after the inclosing strata had been formed; it is therefore a sill. If a sheet has not baked the stratum overlying it, it is probably a lava flow, and at any rate it is older than the overlying strata. Thus the age relationship of an igneous body with the rock masses associated with it is determined by examining its contacts with them.

In recent years much research has been done on methods of determining the absolute ages of minerals and rocks, that is, their ages in years. Certain radioactive elements, notably uranium and thorium, are spontaneously disintegrating at constant rates, and lead is formed as the end-product of this disintegration and accumulates within the parent minerals (p. 26). By the methods based on radioactivity, it was determined in 1938 that the age of the uranium-bearing veins in Joachimstal (or Jachimov), Bohemia, from which Mme. Curie first isolated radium, is 230 million years. Determinations on other minerals from other localities show that the most ancient rocks of our planet are more than 2000 million years old.

TEXTURE AND COMPOSITION OF IGNEOUS ROCK

Texture. The most obvious thing about an igneous rock is its *texture*. By texture is meant the appearance of a rock as determined by the size, shape, and arrangement of its constituents (see also pp. 561–563). Most rocks are made up of mineral grains, but some consist of glass, and some of glass and mineral grains.

If the mineral grains are large enough to be recognized by the unaided eye, the rock is said to be *phanerocrystalline* (Greek *phaneros,* visible, manifest), which can be shortened to *phaneric* (făn-ē′rĭc). If the grains are so minute that they can not be perceived as such by the unaided eye and consequently the rock seems to be a single homogeneous substance, it is said to be *aphanitic* (ăph-a-nĭt′ĭc), meaning "not phaneric."

The grain size is coarser *the more slowly a magma has cooled.* If a magma is extremely hot, the minerals dissolved in it can not crystallize out from it; that is, the atoms and atomic groups in the magma are unable to arrange themselves to form organized solid compounds (the minerals). When the temperature has fallen far enough, the minerals begin to separate from the magma. If the cooling is slow and if the magma has not become highly viscous as a result of the cooling (a very important proviso) they have time to grow to large size, thus forming a coarse-grained rock. But if the magma cools rapidly, more and more centers of crystallization form spontaneously, and, instead of a few such crystal nuclei growing, many begin to grow simultaneously. None of the crystals can therefore grow to a large size, and consequently the resulting rock is fine grained. If cooling is still more rapid, the crystals are so minute that they are invisible to the unaided eye, and the resulting rock is aphanitic. Under conditions of extremely rapid cooling a magma solidifies into a homogeneous substance before any mineral can crystallize. In this event the product is a *glass*, sometimes called a *natural glass.*

To sum up, igneous rocks are *coarse grained, fine grained, aphanitic,* or *glassy,* the grain size being coarser the more slowly the magma cooled.

Porphyry: Porphyritic Texture. So far it has been tacitly assumed that all the mineral grains in a given rock are of uniform size, that is, that the rock is *equigranular* (Fig 360). Many rocks, however, are composed of grains of two conspicuously contrasting sizes: in part of large crystals and in part of much smaller grains, which form a matrix inclosing the larger crystals. An igneous rock of this kind is called a *porphyry* (Fig. 361); it is said to have a *porphyritic texture.* The matrix of a porphyry is the *groundmass,* and the large crystals imbedded in the groundmass are the *phenocrysts* (fē′nō-krĭsts), meaning clearly discernible crystals. The phenocrysts were formed first, and the groundmass somewhat later.

Porphyries are abundant and of many kinds. The groundmasses of the various porphyries have a wide range in grain size: some are medium grained; others fine grained, aphanitic, or glassy. Generally they are aphanitic, as in the lavas, most of which are porphyritic and have aphanitic groundmasses. The phenocrysts also range greatly in size—from those barely perceptible to those several inches in diameter. They differ widely in abundance, ranging from few to very many. In all porphyries, however, the phenocrysts contrast conspicuously in size with the grains that make up the groundmass. This contrast in size is the essential feature of a porphyry. Porphyritic texture is *not* a contrast of colors. Thus a rock made up of light-colored quartz and feldspar, among which are scattered crystals of black mica, all the grains of the three minerals being of about the same size, is *not* a porphyry, although the black mica contrasts strongly with the light-colored minerals.

Texture as Determined by Geologic Mode of Occurrence. Because the texture of an igneous rock depends chiefly on the fluidity[1] of the magma while the minerals are crystallizing out from it, anything in the geologic environment that influences the fluidity of a magma during the critical stage of crystallization must affect the texture of the resulting rock. Rate of cooling is one such factor, for fluidity decreases rapidly as the temperature falls; retention of the gas content is another factor. Obviously, an *intrusive* mass of magma, surrounded and jacketed by the rock masses it has invaded, will lose its heat and gas content much more slowly than an *extrusive* mass, which has issued at the Earth's surface as a lava flow. Consequently the rock of intrusive masses is generally coarse grained in texture, whereas the extrusive rocks as a rule are either aphanitic or glassy.

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[1] Fluidity is the inverse of viscosity.

The volume of an intrusive mass greatly influences the rate of cooling, because a large mass cools more slowly than a small one. Consequently the rocks that make up such large masses as batholiths are coarse grained. The rocks of dikes and sills that were injected into cold rocks tend to be finer grained. The central portion of a thick lava flow may cool slowly enough to become medium grained, whereas a magma that was injected into a narrow fissure in cold rocks is chilled so quickly that it solidifies as an aphanite or even as a glass. Although there are exceptions, the rule holds true that intrusive rocks are coarse or medium grained and extrusive rocks are fine grained to aphanitic.

From the rule just cited, it follows that the coarse-grained rocks, because they were formed within the crust, become visible at the Earth's surface only after they have been stripped by erosion of the rocks that covered them. Probably no granite has been formed at a depth less than 2000 feet below the Earth's surface.

Although, as discussed above, the rate of cooling is an important factor determining the texture of igneous rocks, it is not the only one. Chemical composition also is important; and if, as is reasonable to do, we consider the gas content of a magma as part of its chemical composition, it is probably the single most important factor in determining texture. Under similar conditions of cooling, magmas low in silica and high in iron and magnesia solidify as rocks of coarser grain than do magmas composed of much silica, alumina, and alkalies.

That the texture formed while a magma is solidifying is influenced by geologic environment is vividly shown by the extrusive rocks. These rocks normally consist of phenocrysts imbedded in an aphanitic groundmass. The phenocrysts—the large conspicuous crystals—were formed mainly before the magma was extruded on the Earth's surface, and the fine-grained groundmass was formed after extrusion. The porphyritic texture thus developed records the great change in the environment while the magma was solidifying, whereby the rate of cooling was enormously speeded up and the escape of the dissolved gases was facilitated.

The presence of gases, especially the gaseous water contained in magma, increases the fluidity of the magma or, to say the same thing in another way, decreases its viscosity. The gas content thereby promotes the power of crystallization to an astonishing degree. This enhanced power of crystallization is notably shown by certain dikes associated with intrusive masses of granite. They occur generally as a fringing swarm of dikes around the periphery of the granite masses and were formed soon after the granite. The dikes are made up of large crystals of quartz, feldspar, and mica, individuals several feet in diameter being not uncommon. The extraordinarily coarse igneous rocks of this composition are known as *granite pegmatites*, and from them are obtained the feldspar and plates of mica used commercially. Some granite pegmatites carry radioactive minerals which, as already mentioned, make possible the determination of their absolute ages.

The rock of a volcanic neck is likely to be comparatively coarse grained in spite of the small volume of the neck, because the continuous upward passage of molten material to the surface has heated up the rocks surrounding the conduit. Consequently the last charge of magma that filled the conduit and solidified there when the volcano became extinct cooled slowly and formed a coarse-grained rock.

Composition of Igneous Rocks. The mineral composition of an igneous rock depends on the chemical composition of the magma from which it solidified. As already emphasized, a magma is made up of two parts: a volatile part, consisting chiefly of gaseous water, with carbon dioxide, sulphur, and other substances, amounting at most to a few per cent, and a non-volatile part, consisting of the fixed constituents, chiefly molten silicates. The volatile part is highly important in rock formation because it increases the fluidity of the magma, but it is almost completely eliminated during solidification. The molten silicates crystallize from the magma to form the rocks.

Many thousands of igneous rocks from all parts of the world have been chemically analyzed. The analyses are reported by the chemist in terms of oxides, and accordingly we conventionally speak of the chemical composition of rocks as if the rocks were actually composed of these oxides. For example, when we say that a rock is high in silica we really mean that its chemical analysis shows that much silica is present; but this mode of expression implies nothing as to how the silica is combined in the rock. Actually, the silica may be present in the rock partly in the uncombined state—"free," in the form of the mineral quartz—and partly in the combined state in combination with the metallic oxides as silicates.[1]

The following list gives the ranges in the amounts of the oxides reported for the commoner rocks:

Silica (SiO_2), invariably present, ranges from 40 to 80 per cent.
Alumina (Al_2O_3) ranges from 0 to 25 per cent.
Oxides of iron (generally both FeO and Fe_2O_3), 0 to 20 per cent.
Magnesia (MgO), 0 to 45 per cent.
Lime (CaO), 0 to 20 per cent.
Soda (Na_2O), 0 to 16 per cent.
Potash (K_2O), 0 to 12 per cent.

It will be noticed that only one acid-forming oxide (silica) is present. The other oxides are oxides of the six metals (aluminum, iron, magnesium, calcium, sodium, and potassium) and are in general bases. Oxides of other elements occur in small or minute quantities in rocks but are so much less abundant that they can here be neglected.

Silicic Rocks and Basic Rocks. Silica predominates in all the commoner igneous rocks. Metallic oxides occur in all, but the particular metallic oxides present range from almost nothing to large amounts. A rough general rule governs the composition of igneous rocks. Although the rocks form an unbroken series whose silica content ranges continuously from 80 per cent to 40 per cent, they can be divided into two classes: one in which the silica content is high and the alkali-metal oxides—soda and potash—predominate among the metallic oxides; and the other in which, conversely, the silica content is relatively low, and lime, iron oxides, and magnesia predominate among the metallic oxides. Rocks high in silica are termed *silicic.* Rocks low in silica are termed *subsilicic,* or more commonly *basic,* because they contain abundantly the bases lime, iron, and magnesia. Rocks intermediate between these two classes can be called *intermediate* or *mesosilicic.*

Silicic rocks as a rule are light colored, whereas basic rocks are dark or black, and heavy because of their abundant iron-bearing minerals. Silicic rocks are sometimes termed *acidic,* because the acid-forming radical silica (SiO_2) predominates in them; the *basic* rocks, because of their high content of bases, are relatively low in silica, carrying about 50 per cent.

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[1] In recent years this distinction between "free" and "combined" silica has become of great legal importance in compensation suits brought by workers suffering from silicosis, a lethal disease of the lungs caused by inhalation of siliceous dusts, generally quartz. The destructiveness of dusts to the lung tissues appears to be directly proportional to the amount of free silica in the dusts.

Crystallization of Minerals from Magmas. If a liquid containing a salt, zinc sulphate for example, is boiled down and concentrated to a certain point, not all the zinc sulphate can remain dissolved, and it begins to separate from the solution as a solid in the form of crystals. If the hot solution is allowed to cool, more crystals of the salt are formed, since hot solutions generally can contain more salt than cold ones. A magma also is a solution; it contains dissolved in it various compounds in a more or less unorganized state. If the magma cools slowly enough, the dissolved matter in it separates from it as crystalline grains, the minerals. This crystallization proceeds as the temperature falls until the whole magma has turned into a solid mass of mineral grains. The molten liquid has become *rock.* The minerals begin to separate from any given magma in a definite order, which is governed by their solubility in that magma.

Minerals Common in Igneous Rocks. The minerals that make up most of the igneous rocks are the following:

LIGHT-COLORED GROUP (SIALIC * GROUP)	DARK-COLORED GROUP (FERROMAGNESIAN GROUP)
Orthoclase feldspar	Biotite (black mica)
Plagioclase feldspar	Pyroxene
	Hornblende (an amphibole)
	Olivine
Quartz	Magnetite

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* *Sial* is a combination of the chemical symbols for silicon (Si) and aluminum (Al), the predominant elements combined with oxygen in rocks of this group.

Feldspars, quartz, pyroxene, hornblende, and *biotite* are by far the most common minerals in igneous rocks, and therefore they should be carefully noted. Details regarding them are given in Appendix A.

Since magmas differ in chemical composition, not only the minerals separating from them but also the relative amounts of these minerals must differ. Thus, a silicic magma, containing much silica and subordinate alkalies and alumina, forms a rock that consists mostly of feldspars and quartz; whereas a basic magma, in which lime, magnesia, and iron are abundant, makes a rock containing abundant pyroxene, hornblende, and other *ferromagnesian* minerals, so called in allusion to the iron and magnesium in them.

CLASSIFICATION OF IGNEOUS ROCKS

Basis of Classification. Igneous rocks are of many kinds and varieties. In refined classifications several hundred kinds are recognized, but fortunately the crust of the Earth is made up chiefly of a few common kinds, not exceeding half a dozen. The diversity of igneous rocks is partly the result of the diversity of the magmas that have been and are being generated within the crust and partly the result of the diverse physical conditions under which these magmas have solidified.

Since igneous rocks vary both in texture and in composition, these two variables can be used to classify them. By taking texture as the first criterion for setting the igneous rocks in order, we obtain at once five major classes.

The major classes thus obtained are then subdivided on the basis of their mineral composition—on the kinds of minerals present and the proportions in which these minerals occur. To these subdivisions the actual rock names are given For example, a rock of equigranular texture, coarse enough so that the minerals can be recognized by the unaided eye, and composed of feldspar, quartz, and some biotite, is called *granite*.

By applying these principles, the classification of igneous rocks is obtained which is shown in the table on page 565.

In general, intrusive rocks, especially those occurring in batholiths, are coarse grained and equigranular. Such rocks have long been called *plutonic* rocks, in fanciful reference to Pluto's realm, thought to have been deep within the Earth. They have formed as the result of slow cooling, and the pressure of the overlying rocks caused the magma to retain its gas content until a late stage of solidification, thereby keeping it in a fluid state. Crystallization and the development of a coarse-grained texture was thus promoted.

Extrusive rocks, on the other hand, have been formed by drastic chilling of the magma and the loss of its gas content at the Earth's surface and consequently are characterized by either fine-grained or aphanitic groundmasses. Drastically chilled magmas and those which abruptly lose most of their gas content while cooling are likely to solidify to glasses.

Thus a high-silica magma slowly cooling and retaining its gas content solidifies to a granite. The same magma, however, if erupted at the Earth's surface, will solidify to form a rhyolite, a rock vastly different in texture and appearance from a granite. Rhyolite is therefore the extrusive equivalent of granite.

A magma of medium-silica content solidifies at depth as diorite, but if extruded at the surface it solidifies as andesite. Andesite is therefore the extrusive equivalent of diorite.

A magma of low-silica content—a basic magma of about 50 per cent silica—solidifies at depth as gabbro, but the same magma if erupted commonly solidifies as basalt. Basalt, then, is an extrusive equivalent of gabbro. However, because basic magmas crystallize extremely readily, especially in thicker flows, some extrusions of basic magma solidify as dolerite so coarse in grain as to verge on gabbro.

Diorite and gabbro, although abundant as intrusive masses, do not commonly occur in extensive batholiths as the granites do. They are more common as stocks, sills, and dikes.

Problems of Igneous Geology

The igneous rocks, and their mode of occurrence in the crust as well as on the crust, present many profound problems. The origin of the various magmas, the source of the heat to make them molten, and the cause of their rise from the depths are fundamental problems. Some of the most suggestive evidence is furnished by the magmas that reach the Earth's surface and there give rise to volcanic phenomena; and as volcanism is an important part of igneous geology, these questions are considered in the following chapter.

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CHAPTER 14

VOLCANOES AND VOLCANISM

VOLCANOES

General Description. A volcano is generally thought of as a steep conical mountain having a crater at its top, from which at intervals it ejects gases, rock fragments, bombs, and lava flows. This concept fits Vesuvius, near Naples in Italy, and as Vesuvius is the world's best-known volcano this picture has popularly come to be regarded as typical of all volcanoes. A worldwide survey shows, however, that many of them diverge greatly from this picture. The essential feature of a volcano is an eruptive apparatus, chiefly a conduit that connects a reservoir of molten rock matter in the depths of the crust with the exterior of the Earth. This conduit is referred to as the throat of the volcano, or as the volcanic pipe or chimney. The more or less conical structure that surrounds the orifice of the conduit is called the *volcanic edifice.* The term volcano is customarily used, however, to include both the vent and the hill or mountain built around the vent.

Volcanoes range in size from small cones hardly larger than a beehive to majestic peaks that rank with the loftiest mountains on the globe. Many of the highest peaks of the Andes are volcanoes; some are still active, as Cotopaxi in Ecuador, which has the distinction of being the world's highest active volcano, 19,600 feet in altitude. Cotopaxi erupts from its summit, is unscarred by erosion, and is still in its prime; but many other volcanoes of the Andes are dormant or have become extinct in the recent geologic past. All the Andean volcanoes stand upon an eroded platform of much older rocks, above which they tower 10,000 to 12,000 feet. Certain volcanoes in mid-ocean are of still greater size. For example, the Hawaiian volcanoes stand 30,000 feet above the floor of the Pacific Ocean; their highest summits project 14,000 feet above sealevel.

The higher peaks of the Cascade Range, beginning on the north with Baker, Rainier, and Adams in Washington and extending southward to include Hood in Oregon and Shasta in northern California, are volcanoes that are slumbering or have recently become extinct. Lassen Peak, in California, at the south end of the Cascade Range, is the only active volcano within continental United States.

EJECTION OF MATERIAL

Materials of three kinds are ejected from volcanic vents: gases, liquids consisting of molten rock matter, and solid material in the form of fragments.

Gases. Steam is discharged by active volcanoes in immense quantity, as indicated by the enormous volume of the clouds that accompany many eruptions. These clouds consist of the dust and ash borne aloft by the uprushing column of steam and other gases. The great quantity of steam thus discharged into the atmosphere condenses and causes torrential rains near the volcano; but part of the downpour is probably due to condensation from moisture-laden air that was sucked up to high altitudes by the up-draft created by the discharge of the volcanic gases. Owing perhaps to the friction of the ash particles and to atmospheric disturbance, the eruptions and rains are accompanied by electrical manifestations, which during violent outbursts culminate in spectacular lightning discharges.

The composition of the gases emitted during an actual volcanic eruption is not directly known, because of the difficulty of studying a volcano in action and of capturing its gases uncontaminated by atmospheric gases. The composition differs at different volcanoes and probably at different stages of an eruption. However, indirect evidence shows that the chief constituent is steam. As an example of the immense quantity of steam emitted during the height of an eruption, Paricutin, the world's youngest volcano (Fig. 216), is estimated to have discharged in May, 1945, as much as 16,000 tons a day, along with 100,000 tons of lava.

Besides steam, volcanoes exhale many other gases and volatile products. The emanations are given off not only from the vent itself, but also from the flows of lava, which, as they cool and harden, continue to emit gases for weeks and months. Carbon dioxide, hydrochloric acid, hydrofluoric acid, and hydrogen are given off. To the mixture of the hydrogen with oxygen and its sudden combustion are sometimes ascribed the explosions in the conduit. Sublimed sulphur and compounds of sulphur, such as hydrogen sulphide (H_2S) and sulphur dioxide (SO_2), are emitted by some, but not all, volcanoes.

Chlorides are given off copiously at many volcanoes. In fact, the abundant chlorides at the Italian volcanoes suggested the idea that eruptions are caused by sea water leaking into magma at depth. Even were this inference true for the Italian volcanoes, it can not hold universally, because no chlorides are given off by Kilauea, which is on the ocean-girt island of Hawaii.

Fragmental Products. Fragmental ejecta are the materials blown into the air by explosions in the vent of a volcano. They are derived from the crust of hardened lava left in the upper part of the conduit after a previous eruption; from rock fragments torn from the walls of the conduit; or from new lava ejected from the top of the liquid column of magma in the crater by violently escaping gases. Although clots of lava begin their aerial flight in a liquid condition, they generally harden and fall as solid fragments.

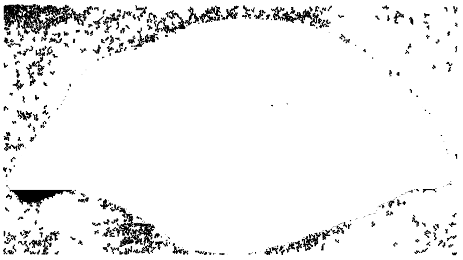


FIG. 200. Volcanic bomb, 6 inches long, with spirally twisted ends. Hawaii.

The pieces of rock and the clots of molten lava blown out and solidified range greatly in size: from dust so fine that it floats in the atmosphere for several years to large masses weighing many tons. The projectile power of volcanoes is astounding. Thirty-ton blocks were blown out by Stromboli in 1930 to a distance of 2 miles; and other volcanoes have hurled smaller blocks to much greater distances.

The fragmental material blown out from a volcanic vent is collectively called *pyroclastic material.* For the sake of precision it is classified according to size, somewhat arbitrarily, as follows: pieces larger than 32 millimeters (1¼ inches) in diameter are called *blocks* if ejected as solid fragments and hence angular, and *bombs* if ejected as particles of still-fluid magma and hence roundish or ellipsoidal; those between 32 millimeters and 4 millimeters are termed *lapilli* (Latin for *little stones*) or *cinders;* those between 4 millimeters and ¼ millimeter are *ash,* and the still finer material is called *fine ash,* or *dust.* Although "ash" and "cinders" are commonly used in referring to the products of eruptions, volcanic ash and cinders are *not* products of combustion. A volcanic bomb is illustrated in Fig. 200; its spirally twisted ends prove that it was still liquid during its flight.

The bombs, lapilli, and most of the ash fall near the vent and thus help to build up a cone around it. The dust is carried by the prevailing winds long distances, hundreds of miles or more, and is thus spread over an immense area. Huge quantities are discharged during great eruptions, amounting to many millions of tons. Such dust showers are highly destructive to vegetation and even to animal life, but the soil ultimately yielded by them is very fertile.

Liquid Material: Lavas. A volcanic cone has no strength and is easily fissured by the explosions and by the pressure of the lava column in the conduit. Hence the magma does not as a rule flow out over the lip of the crater but issues through fissures in the sides of the cone. It may even happen, especially if the cone is built of cinders, that, unable to withstand the hydrostatic pressure of the lava column in the conduit, one side of the cone gives way and allows the lava to rush out through the breach thus made. Such repeated breaching of the side, giving rise to repeated lava flows, occurred in 1944–1947 during the growth of Paricutin, the newborn volcano in Mexico.

The appearance and character of a lava stream and the rock formed from the lava as it solidifies depend on several things. The chemical composition of the magma determines the nature of the resulting rock, whether it will be a light-colored lava (rhyolite) or a black basalt, or of intermediate character (andesite), as previously explained. The fluidity determines the rate at which the lava flows, the distance it flows, and in large measure the aspect of its surface. When the lava first flows out, it is red or white hot and highly fluid. It soon cools on the surface, darkens, and crusts over. As it cools it becomes more and more viscous. When the flow becomes very viscous, the under part may still be moving while the upper part crusts over and breaks up into rough, angular, jagged blocks, which are borne as a tumbling, jostling mass on the surface of the slowly moving flow. When eventually the flow comes to rest and solidifies, the resulting lava sheet is extremely rough. Its top is a chaotic assemblage of blocks and scoriaceous fragments, bristling with innumerable sharp points. Such lava flows are termed *block lava.* In Hawaii they are called *aa* (ä'ä').

In marked contrast to the block lava, other flows harden with smooth surfaces, which have curious ropy, curved, and billowy forms, as seen in Fig. 201. "Corded" lava of this kind the Hawaiians term *pahoehoe* (pä-hō'ā-hō'ā). Fine examples of corded lava flows can be seen in the

Craters of the Moon National Monument in southern Idaho. Although these flows are at least several hundred years old, the surfaces still show iridescent colors in blue, purple, and bronze, as if the flows had been erupted only a few years ago.

Very fluid lavas flow rapidly, especially on steep slopes (Fig. 3). Some, like the great flow from Mauna Loa in 1850, average 10 miles an hour. However, speeds of more than 5 miles an hour are exceptional. As the lava flows cool and become viscous, they move extremely slowly, creeping onward, possibly for several years.

W. C. Mendenhall, U. S. Geological Survey.

FIG. 201. Lava cascade, Hawaii. Basaltic magma is solidifying into corded lava called pahoehoe.

After a lava flow has crusted over, the still-liquid portion in the interior may burst through the lower end of the flow and run out, draining the interior. A long tube called a *lava tunnel* is thus formed. The natural drainage of some volcanoes passes into these tunnels, disappears from view, and issues lower on the slopes as springs.

In 1935 lava broke out on the flank of Mauna Loa at an altitude of 9000 feet and flowed toward the town of Hilo, threatening to destroy it and its harbor, the second most valuable in the Hawaiian Islands. The U. S. Army Air Force then tried bombing the flow in order to halt it and avert the threatened disaster; the attempt was reported to have been successful. Success depended on the fact that the lava was flowing through a lava tunnel toward Hilo. The bombing demolished the roof of the tunnel at the upper end of the flow and blocked the tube with debris. The lava was thereby forced to overflow, cool, and solidify; and the danger was averted.

Effects of Escaping Gases: Vesiculation of the Lava. That lavas, even after they have issued from the volcanic vent, still contain dissolved gases is amply shown not only by the clouds of steam that escape from them for weeks and months but also by the structures they assume as they solidify into rock. Viscous lava may become highly inflated by the expansion of innumerable bubbles of gas. Each bubble hole is a vesicle, which is spherical if the lava was stationary and almond shaped if the lava was moving and drawing out the vesicle while the hole was forming. The upper portion of a flow, especially of a viscous silicic lava, may contain so many bubble holes that it has become a veritable froth. Rock froth, which is generally white or light colored, is known as *pumice.*

In more fluid lava, especially basalt, the gas cavities, or vesicles, attain larger sizes. If the cavities are highly irregular in shape and size and are so abundant that there is at least as much empty space as solid matter, the resulting rock is *scoriaceous.* Loose pieces of such material are called *scoriae;* they are commonly dark, black, or reddish.

Phases of Volcanic Activity. We have seen that volcanoes eject materials of three kinds: gases, liquid lava, and solid fragments. The nature of a volcanic eruption depends largely on how much of each is ejected. According to its phase of activity, a volcano emits very different proportions of these products. Volcanologists generally characterize the activity a volcano is displaying at a particular stage in an eruption as hawaiian, strombolian, vulcanian, or peléan, terms indicative of the most characteristic habits of certain well-known volcanoes.

Hawaiian Phase of Activity. In the mildest phase of activity volcanoes tranquilly discharge flows of lava, unaccompanied by the explosive escape of gases and the ejection of fragmental material. The lava thus discharged is very hot and highly fluid. More or less gas is given off, but it is released without catastrophic violence. Well-known examples are in Hawaii. Therefore this nonexplosive phase of activity is termed hawaiian.

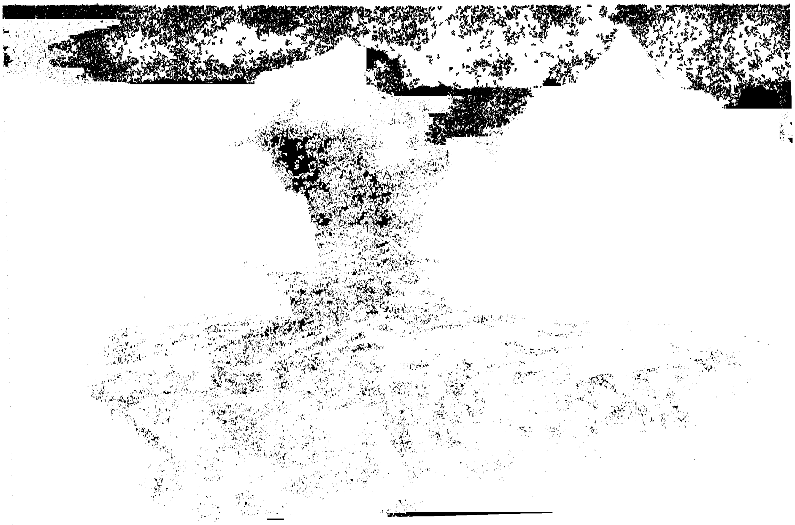
The island of Hawaii is built of an immense mass of basaltic lavas crowned by five great volcanoes, of which Mauna Loa (13,680 feet high) and Kilauea (4000 feet high) are active. Mauna Loa exceeds all other active volcanoes in its vast size and in the amount of lava it erupts. Immense columns of white-hot lava play as fire fountains at times 700 feet high, and afford truly spectacular displays. Its flows of lava generally issue from its flanks rather than from the summit crater, and sometimes they issue below sealevel.

On the eastern slope of Mauna Loa, 22 miles from its top and 10,000 feet lower, is Kilauea. In the summit of Kilauea is a depressed area, rudely oval and 9 miles in circumference, shut in except on the south by vertical cliffs, which are as much as 450 feet high. In the floor of this depression is Halemaumau, the "Everlasting House of Fire," a pit which before 1924 was 1300 feet in diameter and contained a lake of liquid basalt, "boiling" and fountaining because of the abundantly escaping gases (Fig. 202). The "pit" is the upper end of the conduit of Kilauea. In 1924 the column of lava in the pit suddenly dropped 700 feet, and this subsidence was followed by explosive eruptions, the first since 1790.

The explosions of 1924; after an uninterrupted period of 134 years of quiet extrusion of magma, illustrate the principle that even basaltic volcanoes, which usually extrude the most fluent of lavas, become at times violently explosive. As the result of the sinking of the lava column to great depth in the volcanic pipe, the ground water in the

HAWAIIAN VOLCANO OBSERVATORY.

Fig. 202. Halemaumau, the fire pit of Kilauea. October, 1921.



surrounding edifice suddenly entered into the pipe. There it encountered hot rock and magma, generated immense quantities of steam, and caused the spectacular explosive eruptions. Enormous cauliflower clouds were emitted, and large amounts of old rock that had avalanched into the pipe were blown out.

Strombolian Phase. The volcano Stromboli is in the Mediterranean Sea north of Sicily, in the Lipari or Aeolian Islands. It has been continuously active since the dawn of history, its eruptions as a rule consisting of mild explosions every 10 or 15 minutes, whereby clots of fluid or viscous lava are hurled into the air and fall back into the crater. Two features of this eruptive activity are notable: first, the magma does not freeze over between successive explosions; and, second, despite the great amount of heat continuously lost from the open crater, this loss has been made good by supplies coming from the depths, probably by up-streaming gases, during the past 2000 years. Stromboli ejects hot, incandescent material, producing luminous eruption clouds; hence Stromboli has been called the lighthouse of the Mediterranean. A volcano, accordingly, is said to be in a strombolian state when it blows out clots of incandescent lava and the magma in its crater does not crust over between consecutive explosions.

Vulcanian Phase. Vulcano is another active volcano in the Lipari Islands. Its magma is stiffly viscous and crusts over solidly between consecutive explosions. This crust is shattered during the next eruption, and angular fragments of rock are therefore blown out by explosions of this kind. The eruption clouds rise vertically from the crater, are dense, and take on cauliflower-like forms, but even at night they are dark; that is, they are not incandescent.

Ultra-vulcanian is a term often used for eruptions in which only old, cold material is blown out. Activity of the ultra-vulcanian type is most likely to occur during the clearing out of a volcanic vent after the volcano has been dormant for some time; and it is known to take place regardless of the kind of magma a volcano emits, whether silicic, intermediate, or subsilicic (p. 299). The explosive eruption of Kilauea in 1924 just described was an ultra-vulcanian eruption from a volcano whose prevailing activity is of the hawaiian type.

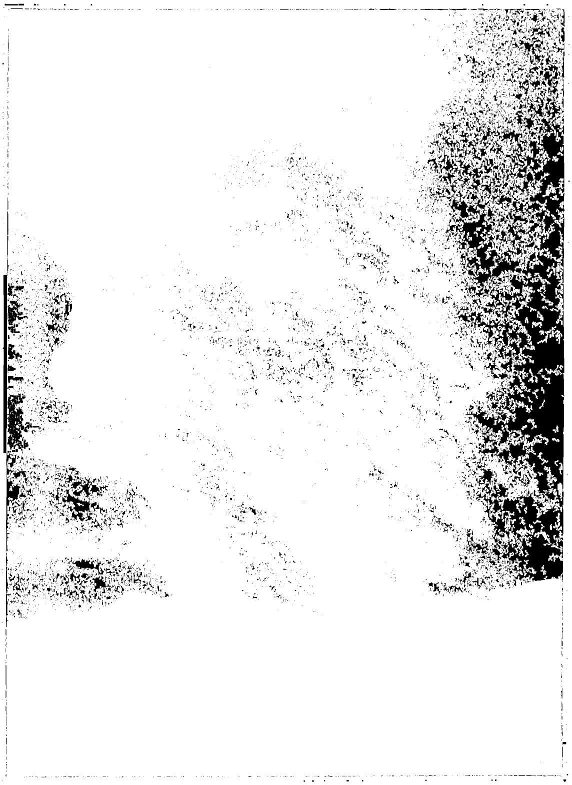
Peléan Phase. The peléan phase is the most violently explosive of all, and its eruption clouds are denser even than the vulcanian.

The volcano of Mont Pelée on the island of Martinique in the West Indies, after mild premonitory symptoms, began in May, 1902, to erupt in a series of violent explosions. It had been in repose since 1851, and at that time had been only feebly eruptive. The probability therefore is that the internal pressure then existing had been only partly relieved and that during the long dormant period—51 years—an immense gas pressure was built up. No lava was discharged during the eruption of 1902, but, on the morning of May 8, an enormous dense black cloud of superheated steam heavily laden with fragments of extremely hot gas-charged lava blew out horizontally from the flank of the mountain and sped down the slopes to the sea, at a rate of more than a mile a minute. Annihilating all life in its course, the cloud swept through the town of St. Pierre 5 miles from its point of origin and instantaneously destroyed the town and its 26,000 inhabitants and the 2000 who in the previous days had fled to it for safety. Only two persons, one of them a prisoner in the dungeon, are known to have escaped alive. Almost at the same time La Soufrière on St. Vincent, 90 miles away, ejected a similar destroying blast, the "Great Black Cloud," which killed 2000 people and devastated a broad tract of country. For many months after, Mont Pelée continued at irregular intervals to eject these clouds, one of which is shown in Fig. 203 as it arrived at the sea. During 1929–1932 there was a mild repetition of the activity so characteristic of 1902. This relative mildness is ascribable to the fact that the interval had been too short to build up the tremendous gas pressure that obtained in 1902.

The great eruption of Pelée in 1902 directed attention to a phenomenon previously unknown. The French investigators called the remarkable clouds *nuées ardentes,* meaning scorching or glowing clouds. These remarkable clouds are now called *peléan clouds.* The extraordinary features of a peléan cloud are: it is ejected horizontally from beneath the lava plug in the summit of the volcano instead of blasting the plug out vertically upward; it is extremely hot; it travels at great speed; it carries with it an enormous quantity of highly gas-charged lava fragments, including blocks many yards in diameter. All this material is deposited chaotically together, instead of raining down according to size, as from other volcanic clouds. In reality, the cloud hides the vital feature of the ejection—the immense quantity of hot gas-emitting lava fragments moving close to the ground at enormous speed, more than a mile a minute.

Peléan clouds having drawn attention to themselves so forcibly as features of volcanism, the question arose whether other volcanoes have ejected them in the past. If so, by what criteria can such eruptions be recognized? Later we shall see that peléan clouds are regarded as having been ejected in immense numbers and in unprecedented magnitude from the volcano that once stood at Crater Lake, Oregon.

Factors Determining Explosive Power of Magmas. The explosive power of a magma depends primarily on how much gas it contains. If it contains no gas, it has no explosive power. Furthermore, viscosity influences the readiness with which gases are liberated from magmas.



A. Lacroix.

FIG. 203. Cloud from Mont Pelée, Dec. 16, 1902, arriving at the sea. Height of cloud is 13,000 feet.

Viscosity in turn is enormously influenced by the composition and by the temperature of the magma, being very much less at higher temperatures. Because temperature and gas content of a magma can change, the explosive tendency of the magma can change; consequently, at different times the same volcano may go through several different phases of explosive activity.

Explosive activity is much more common in volcanoes that eject andesitic and rhyolitic lavas than in basaltic volcanoes. The greater explosive power of andesitic and rhyolitic volcanoes in comparison with that of basaltic volcanoes is undoubtedly the result of the higher gas content of their magmas.

VARIETIES OF VOLCANOES AND CRATERS

Pyroclastic Cones. The shape and structure of the edifice built up around a volcanic vent depends on the material of which it is formed. If the edifice is made wholly of fragmental ejecta, a steep cone is built. Slopes of 30 degrees, rarely 40 degrees, are attained before the accumulating mass begins to slide. Volcanic edifices of this kind are called *pyroclastic cones;* they are characteristic results of explosive eruptions. A pyroclastic cone built of huge angular blocks forms the summit of Etna (p. 318). If a pyroclastic cone consists wholly of cinders and lapilli, it is called a *cinder cone.* Cinder cones are relatively small; none of them exceeds a height of 1000 feet (Fig. 204).

Shield Volcanoes. Many volcanoes have been built largely or entirely by the piling up of lava flows one upon another around a central

SPENCE AIR PHOTOS.

Fig. 204. Cinder cone, 200 feet high. Amboy, California.

vent. A volcano built in this way is low in proportion to the immense area it covers and shaped like a shield, its summit constituting a gently convex plateau; hence it is called a *shield volcano* (Fig. 205). Volcanoes of this kind are built up by the emission of highly fluid lava flows, which spread widely as thin, nearly horizontal sheets. They are well represented in the Hawaiian Islands. Mauna Loa, 13,680 feet high, is a vast shield volcano, the largest in the world: it is an enormous, gently sloping dome; its slope near its base is 2 degrees, increasing summitward to 10 degrees, but flattening above an altitude of 10,000 feet.

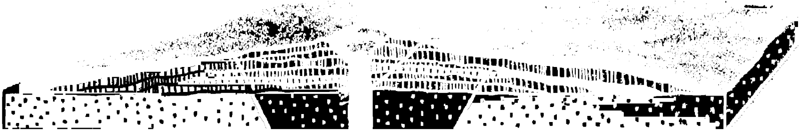


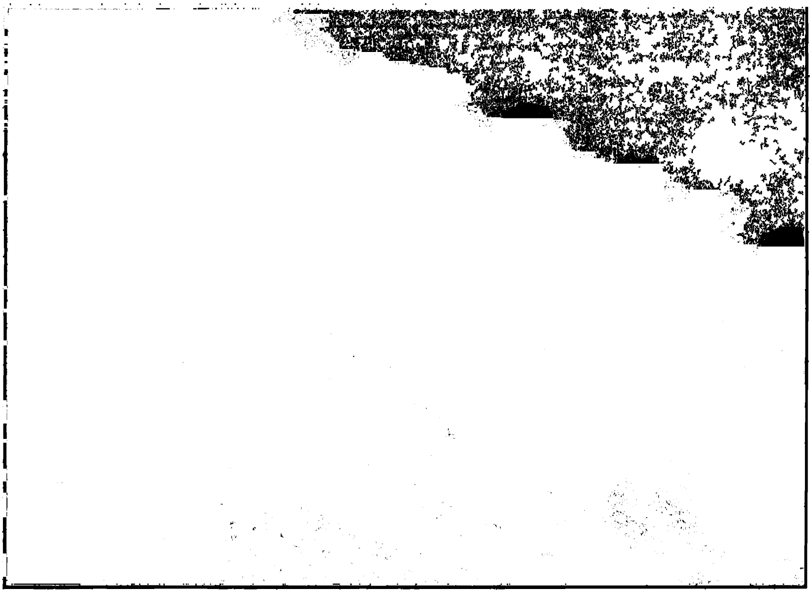
FIG. 205. Section through a shield volcano. Sills, dikes, and the filling of the central conduit are shown in black.

Stratovolcanoes. Most volcanoes, including many of the world's larger volcanoes, are intermediate in form between the steep pyroclastic cones and the flat domes of shield volcanoes. They are built of layers of fragmental ejecta alternating with lava sheets. This composite structure has resulted from the fact that the volcanoes were at times explosively active and at other times were quietly discharging streams of lava. The rude layering, or stratification, thus produced slopes downward and outward from the conduit of the volcano (Fig. 206). Such composite cones are called *stratovolcanoes* (Fig. 207).

The ejecta composing the layers gradually become compacted by their weight and by the infiltration and deposition of cementing substances. As previously mentioned, the coarser material forms volcanic breccia, and the finer material forms tuff.

Since lava rarely flows over the lip of the crater but, especially in high volcanoes, issues from fissures in the sides of the volcano, these fissures also become filled with molten lava, which hardens into rock, forming dikes. The dikes radiate from the vent as a center, and they serve as ribs to strengthen the volcanic cone. Thus a vertical section through a volcano of this kind reveals a central core of massive igneous rock or of vent agglomerate surrounded by beds of tuff and breccia and intercalated sheets of lava, which are cut by a radial system of dikes.

As a result of its composite construction a volcanic cone of this origin has a characteristic profile of great beauty: with concave slopes, gentle near the base steepening upward to the summit crater. Mayon, the most active volcano in the Philippine Islands, is generally regarded as the finest example of a stratovolcano (Fig. 208). Many others the world over have similar graceful contours, but the most famous is Fujiyama, the sacred mountain of Japan, towering 9000 feet above its surroundings. Within our own country are the majestic cones of the Cascade Range, all of them stratovolcanoes—Rainier, Adams, Hood,



W. T. Lee, U. S. Geological Survey

FIG. 206. Part of a volcanic cone, showing inclined beds of tuff and breccia. Trinchera, Colorado.

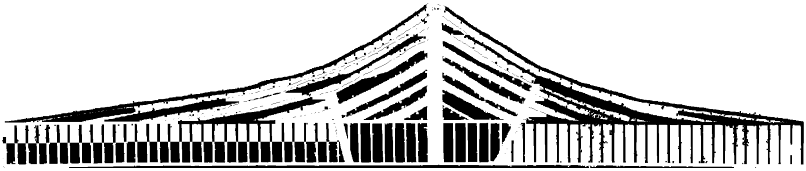


FIG. 207. Stratovolcano. The internal layered structure has resulted from the alternation of breccias, tuffs, and lava sheets dipping away from the central vent. The profile is that of Mayon (Fig. 208).

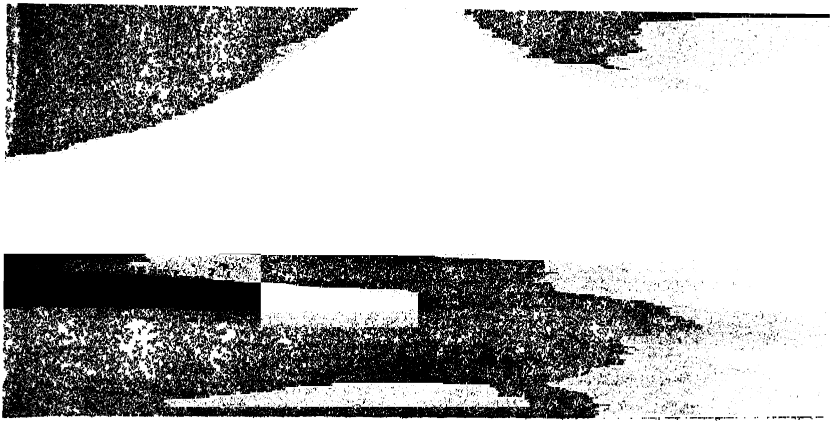


Fig 208. Mayon, Philippine Islands; height 7900 feet. One of the world's finest examples of a stratovolcano.

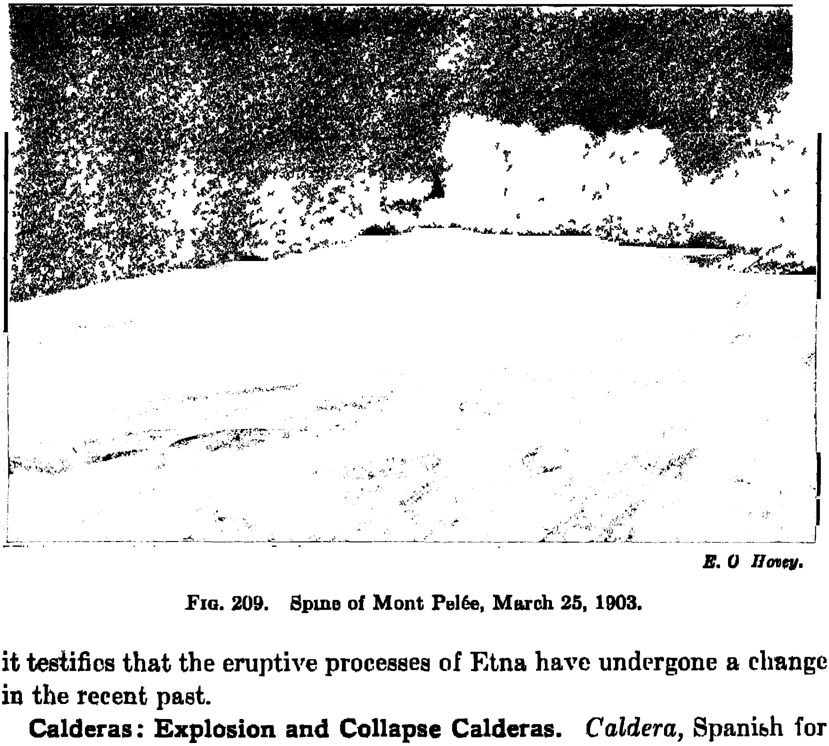
Shasta, to name only some—which give the range its distinctive scenic grandeur.

Plug Domes. At the time it is being extruded, a magma may be too viscous to flow readily. Consequently the sluggish pasty mass piles up over the vent as a great dome, termed a *plug dome.* A plug dome grows by expansion from within. As it becomes larger, the exterior of the dome hardens and forms a carapace, but, because of the continuing expansion of the dome, the carapace cracks and sheds blocks, which roll down the flanks and form immense accumulations of angular blocks. Such doming is confined chiefly, though not wholly, to silicic lavas.

Thirteen domes formed in this way occur within an area of 50 square miles in Lassen Volcanic National Park, the largest being Lassen Peak itself.

After the violent eruption of Pelée in 1902, the column of silicic magma that filled its vent hardened into rock at the surface but remained highly viscous in the interior, and the whole mass was pushed up so that it rose like a vast tower above the volcano and eventually, a year after the destruction of St. Pierre, attained a maximum height of 1200 feet above the crater rim. As viewed from the northeast, the tower resembled a gigantic spine and was therefore called the spine of Mont Pelée (Fig. 209). Gradually it crumbled into a mass of blocks as a result of continuous explosion of gases.

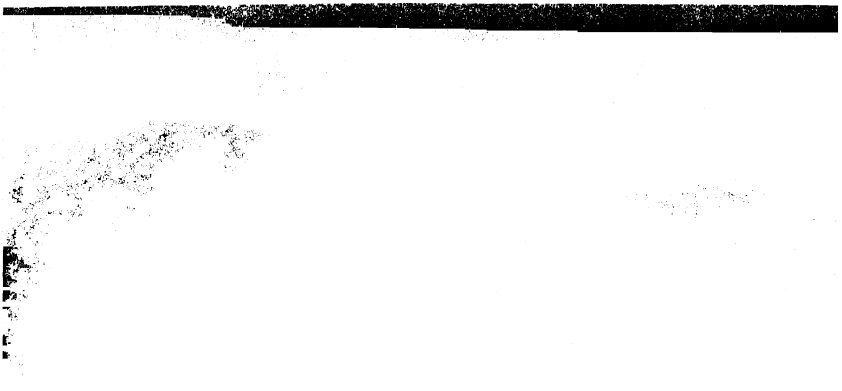
Compound Volcanoes. Pyroclastic cones, stratovolcanoes, and shield volcanoes represent only the simpler types of structures. Many existing volcanoes are actually much more complex, and are therefore called *compound volcanoes.* Etna, which Sicilian geologists proudly call "the most classic volcano in the world," is a good example of a compound volcano. The main bulk is made up of lava flows, constituting a vast shield volcano. On its summit stands a pyroclastic cone, made up of huge angular blocks. It is 1000 feet high and is thus small in comparison with the great bulk of Etna, but it is of great interest as it testifies that the eruptive processes of Etna have undergone a change in the recent past.



E. O. Hovey.

FIG. 209. Spine of Mont Pelée, March 25, 1903.

Calderas: Explosion and Collapse Calderas. *Caldera,* Spanish for caldron, is the term applied to crater-like basins of volcanic origin that are of great size, especially those which are wide compared to their depth. The term was taken from the huge circular pit in the Canary Islands, called La Caldera, which is 3 to 4 miles wide and is bounded by lofty cliffs 1500 to 2500 feet high, except on one side where the encircling wall has been breached by erosion. As seen from a distance, the mountain inclosing a caldera resembles a huge cone that has been truncated far below its apex. As viewed from its floor, a caldera is a wide plain ringed by cliffs. In this plain there may be active volcanoes, or



U. S. ARMY AIR CORPS

Fig. 210. Crater Lake, Oregon, a lake in a caldera. The island, Wizard Island, is the summit of a small volcano built after the top of the main volcano had collapsed.

there may be extinct volcanoes, or there may be active and extinct volcanoes together.

Many calderas occur throughout the world Some were formed by gigantic explosions that blew away the tops of former volcanic cones as dust and ash. Great circular pits mark the sites of these wrecked volcanoes. Calderas definitely known to have been formed by explosion rarely exceed 1 mile in diameter.

Calderas, especially those more than 1 mile in diameter, are formed by the collapse and engulfment of the summit portions of volcanoes. The collapse may have been caused by subsidence along a circular fracture, as at Kilauea, or it may have resulted from the rapid emptying of the magma reservoir beneath the volcano.

Crater Lake, in southern Oregon, is at the summit of a volcanic mountain in the Cascade Range. A mysterious place for a lake, thought its astounded discoverers. The lake, justly famous for its beauty and the marvelous blue color of its water, is 5 miles in diameter, 2000 feet deep, and encircled by steep cliffs 500 to 2000 feet high (Fig. 210). These cliffs reveal clearly the anatomy of an old volcano; and the lake therefore manifestly occupies a caldera—not a crater, for its diameter of 5 miles greatly exceeds that of any known eruptive vent. An island in it—Wizard Island—is the unsubmerged top of a small but perfect cinder cone, which shows that feeble eruptions broke out after the caldera had been formed. The caldera, if emptied of the water now in it, would appear as a great circular pit 4000 feet deep.

The volcano that formerly stood here, named Mount Mazama, was built up to a height of 12,000 feet, about 6000 feet above the platform on which it stood; during most of its life it was heavily capped with snow and glaciers. The upper 6000 feet of the volcano have disappeared. The volume of the vanished portion is 17 cubic miles. Where has this material gone? The reason for believing that the caldera was formed by collapse and engulfment of the top of the former volcano, rather than by explosion, lies in the absence of the debris—the immense volume of tuff and breccia—that so gigantic an explosion would have spread over the immediately adjacent outer slopes. Late in the life of the volcano, just before the final catastrophe, there were enormous eruptions of pumice, which culminated in the ejection of *peléan* clouds of extraordinary magnitude and intensity. Some were so powerful that they traveled 34 miles from the volcano. Magma was also drawn off subterraneously from the conduit. Thus the superstructure of the volcano was weakened, and it collapsed into the partly emptied magma reservoir, forming the caldera. This catastrophe is estimated to have happened 10,000 years ago.

There is no sharp demarcation in size between a crater and a caldera. The essential feature of the caldera, as already emphasized, is that its diameter is generally enormously larger than that of the supply pipe of the volcano. However, craters enlarged by a reaming out, as it were, are not calderas. During periods of repose of a volcano, avalanching from the walls into the crater occurs, thereby increasing the upper diameter of the crater. Subsequently the fragmental material that fell into the crater may be blown out. During continuous blow-offs of gas, such as characterized the great eruption of Vesuvius in 1906, avalanching of this kind and the ejection of the avalanched material enlarge the crater many fold, but craters thus reamed out are genetically different from calderas.

Explosion Pits. In some places volcanic activity has gone no further than to drill a vent through the country rock. The material blown out consists therefore largely of fragments of the country rock mixed with only a minor amount of lava fragments. Because of its small volume, the ejected material has made only a low ridge around the pit, but no real cone was built (Fig. 211). Such pits range in width from a few hundred feet to possibly 2 miles. In moist regions they are filled with water and form lakes.

Vesuvius. Vesuvius has by long and careful study become the best-known volcano in the world. It is therefore of interest to present some of the facts regarding this volcano and to give the explanation advanced to account for its eruptive mechanism.

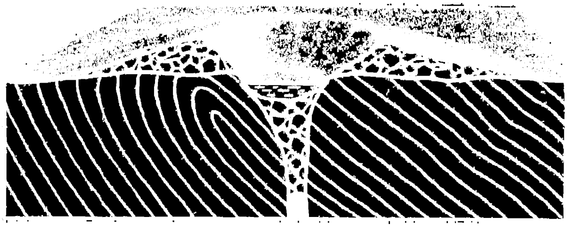


FIG. 211. Explosion pit occupied by a lake. The low cone around the vent consists chiefly of fragments of the country rock perforated by the vent. An explosion pit is an *embryonic* volcano.

Vesuvius stands on the site of an older volcano, Somma. To the Romans Somma seemed to be extinct, for, although they recognized its volcanic nature, they had no traditions of its having been active. In

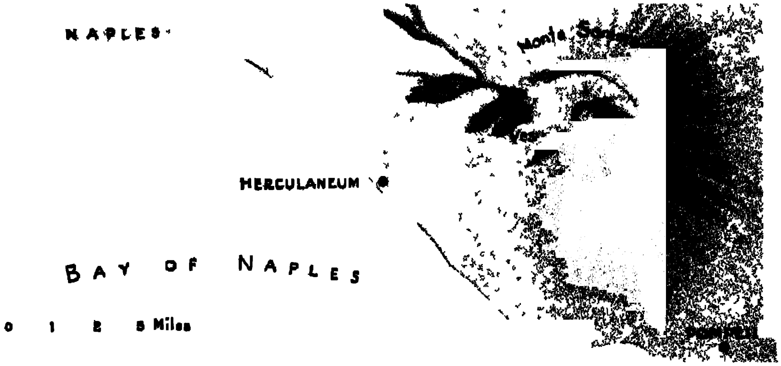


FIG. 212. Map of Vesuvius and vicinity. Some of the larger streams of lava erupted in recent years are shown.

A.D. 79 the volcano became violently active in eruptions that destroyed the towns of Herculaneum and Pompeii on its seaward flanks. Much of Somma on the side toward the sea was blown away or engulfed, and a caldera was formed. In this caldera Vesuvius began to be built and is now 4000 feet high. Partly inclosing it is the crescentic ridge of Monte Somma, the remains of the rim of the caldera (Fig. 212).

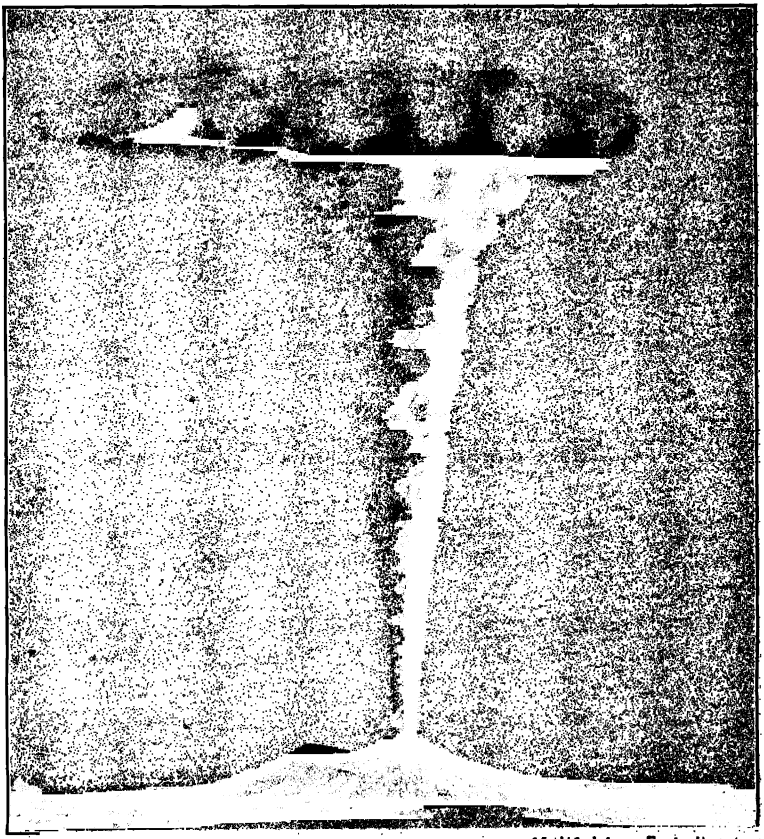
Since 1631, Vesuvius has been in a state of almost constant mild activity, with irregular periods of violent eruption. The three greatest eruptions in its history occurred in 79, 1631, and 1906. In the eruption of 1906 three stages of activity were recognized. During the first stage, lasting four days, incandescent floods of lava were discharged from fissures that opened successively lower and lower on the flank of the cone. Earthquake shocks increased in violence and frequence, so that in effect there was a continuous earthquake for several hours. Great explosions of strombolian type blew immense quantities of incandescent lava thousands of feet into the air. By night the glowing clots illumined the surrounding region as by a gigantic pillar of fire. During the second stage gas was emitted in vast quantity; it escaped under enormous pressure and ascended as a continuous blast with enormous speed to a height of 8 miles, where it spread laterally as cauliflower clouds (Fig. 213). It lasted all of the daylight of April 8, as a continuous blow-off. The discharge of this colossal jet of gas enlarged the crater. This stage was succeeded by the ejection of dark ash expelled by powerful isolated explosions. The whole eruption lasted 18 days, and when it ended the crater was 2200 feet wide at the top and 2000 feet deep.

The mechanism of this great eruption is conceived to be as follows. The crater slowly filled with magma, which gradually by absorbing ground water and oxygen became superheated and gas charged. Finally the magma bored through the walls of the crater; escaping through the perforations and through fissures, it flowed out, thereby diminishing the pressure on the underlying gas-charged magma. The stupendous blow-off of the highly compressed gas then ensued, and the crater was reamed out. After the blow-off ceased, material began to avalanche into the crater. The relatively cold debris that thus fell into the crater was blown out by intermittent explosions, making the final, dark-ash phase of the eruption. Repose then set in, and Vesuvius began to recuperate.

Not till 1913 did the crater begin to fill with lava. At length, in the fall of 1926, the accumulated lava flowed over the top, and again in 1927 and 1928. In 1929 innumerable gas vents gave forth much hydrochloric acid and deposited iron and alkali chlorides. The crater became a lake of brilliantly incandescent lava. Pieces of slag were hurled more than 4 miles. Seismic activity, emanating from a shallow focus, was violent. Lava flows were poured out, extraordinarily thinly fluid because of their large gas content and their high temperature (about

1100°C.).' Descending the mountain, they overwhelmed several small villages.

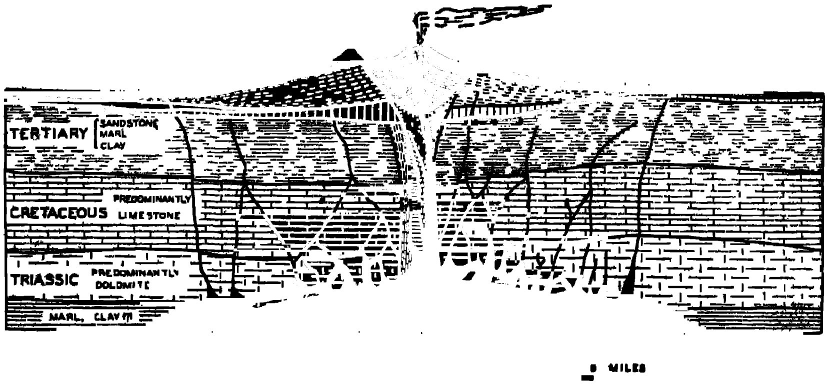
On March 18, 1944, when the Allied troops had reached Naples in their northward advance through Italy, Vesuvius broke out in a spectacular eruption. Many lava flows were discharged, and immense quantities of ash and cinders were blown out. After a week's activity, the eruption ended as suddenly as it had begun. It was the biggest eruption of the century.



Modified from F. A. Perret.

FIG. 213. Climax of the great gas blow-off, during the eruption of Vesuvius in 1906.

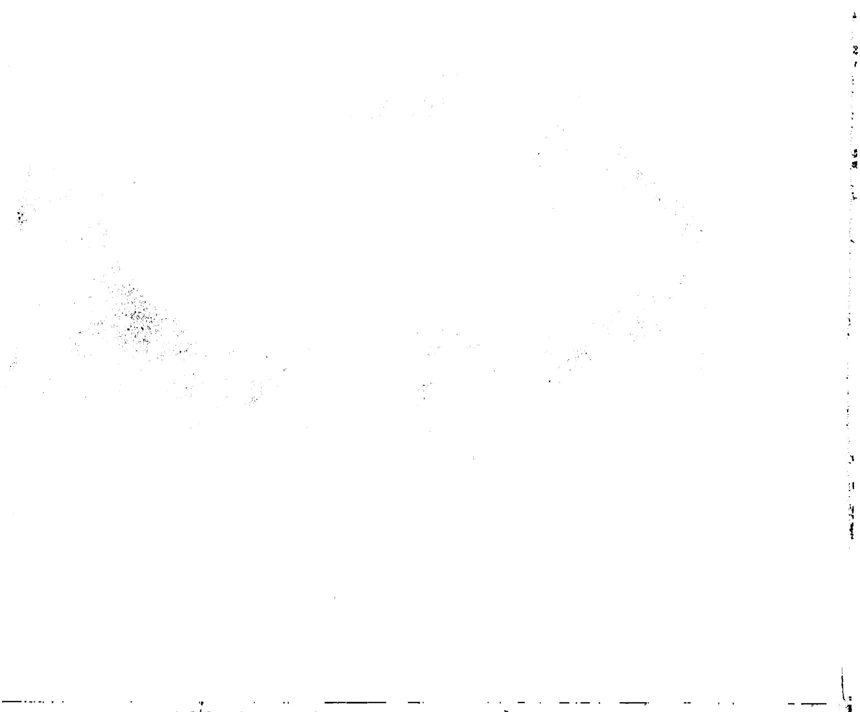
Vesuvius erupts lava of a rare type. This lava differs markedly from the material erupted during the initial outbreak of its ancestor, Somma, about 9500 B.C. Vesuvius and Somma have blown out a great variety of rocks torn from the foundations and the magma reservoir; about 25 different kinds have been recognized. From these ejected blocks, from the surface geology, and from study of earthquake records, the magma reservoir from which Vesuvius draws its lavas and explosive energy is inferred to be at a depth of 15,000 to 18,000 feet (Fig. 214). The magma in this reservoir is slowly solidifying, and the gas content becomes concentrated in the continuously decreasing fluid portion, thereby building up pressure and explosive power. Some evidence, based on the change in chemical composition of the lavas erupted during the life of Somma and its successor, Vesuvius, suggests that the magma has been dissolving the limestone and dolomite wall rocks surrounding the reservoir. It thereby augmented its gas content and its explosive power by the carbon dioxide acquired by dissolving the carbonate rocks. In this respect Vesuvius is an abnormal volcano.



Modified from A. Rittmann.

FIG. 214. Section through Vesuvius, showing the inferred magma reservoir at depth.

Meteorite Craters: "Pseudovolcanic" Features. A pit strongly resembling the explosion pit of an embryonic volcano indents the plateau near Flagstaff, Arizona (Fig. 215). It is nearly circular in outline, is 4000 feet across and nearly 600 feet deep. It is sunk into horizontal sedimentary strata. Encircling it is a low ridge made up of a wildly chaotic assemblage of immense blocks of rock. Fragments were blown out as far as 6 miles from the pit. Some 20 tons of iron meteorite fragments found in and around the crater led to the idea that the crater was formed by the impact of a huge meteorite. Exploration for the meteorite by drilling twenty-seven boreholes and sinking nine shafts has established that the strata beneath the pit are completely intact



N. G. S. PHOTOGRAPH BY ALBERT W. STEVENS; COURTESY NATIONAL GEOGRAPHIC MAGAZINE.

Fig. 215. Meteor Crater, Arizona, as seen from the southeast.

and undisturbed. Thereby the pit has been conclusively proved to be of nonvolcanic origin. This and other cogent evidence leaves little doubt that the pit was formed by the impact of a meteorite and the resulting explosion caused by heat generated by the impact. Hence the pit has been named Meteor Crater.

In recent years many meteorite craters have been recognized in widely separated localities in the world. In the opinion of ballistics experts any meteorite more than 30 feet in diameter would on striking the Earth generate enough heat to volatilize completely. This heat would create at the place of impact a pseudovolcanic explosion.

EROSION OF VOLCANOES

Destruction of the Cones. During all its life a volcano is subject to weathering and erosion. Its height and general appearance at any given time are determined by the balance between these destructive forces and the constructive forces of volcanism. Even an active and growing volcano is trenched by ravines and gulches. After explosive eruptions, when the cone is freshly covered with dust and ash, this loose material becomes so saturated with water from the rainfall that it slides down as flows of liquid mud, forming gullies which become enlarged by subsequent storms. Such a mudflow overwhelmed the city of Herculaneum on the flank of Vesuvius in A.D. 79. It hardened like concrete and covered up the town so thoroughly that Herculaneum became completely forgotten and was discovered in 1738 only by accident. The superior preservation of papyri and works of art in Herculaneum compared to those in Pompeii is due to the effective sealing provided by the concrete-like covering.

As soon as a volcano becomes extinct, the ravages of erosion are no longer repaired, and demolition begins. The tuffs and breccias are carried away rapidly; the harder, more resistant flows and dikes and the parts protected by them are eroded more slowly. It is surprising, however, how long cones that are built of loosely piled cinders resist erosion and retain their form. They resist so well because they are built of porous material, which lets the rainfall sink into it without causing runoff.

As erosion progresses, the volcanic neck consisting of the material filling the supply pipe of the volcano becomes exposed to view. When the cone is demolished, the volcanic neck, because of its greater resistance, generally forms a conspicuous prominence; and, when erosion has finally swept away all external evidence of the volcano and has revealed the foundation rocks on which the volcano stood, the neck remains projecting, a monument to the vanished volcano. Eventually it also vanishes, the "roots" of the volcano become more and more deeply exposed, and finally all evidence that a volcano was once present is obliterated.

Extinct volcanoes occur in every quarter of the globe, many of them where volcanic activity has long ceased. All stages of demolition are represented among them: from cones only slightly attacked by erosion to those so greatly eroded that the original shape has disappeared, but whose central rock shaft (the volcanic neck), outlying concentric masses of lavas, tuffs, and breccias, and radial dikes still plainly show that they were once volcanoes.

The demolition of a volcano by erosion is most easily visualized if the successive stages are shown within the area of a single volcanic field. In such a favorable area, because of the accidents of erosion, some volcanoes have been left nearly intact, others have had their outer framework removed, and still others have been almost wholly destroyed.

Obviously, the internal structure of an active volcano is incompletely shown. Rarely are we given such an opportunity as at Krakatoa, in the Strait of Sunda near Java, where the catastrophic explosion of 1883 blew away half the cone and revealed the anatomy of the volcano. In general we must depend on erosion to reveal the internal structures of volcanoes. In this way, for example, it has been ascertained in certain deeply eroded volcanoes that their supply pipes change into dikes at depth. From this evidence we conclude that some dikes on approaching near the Earth's surface have blown out an orifice.

Some confirmatory evidence has been gathered from actual mining operations. A few volcanic necks contain economically valuable substances: the famous diamond-bearing pipes of South Africa are the most noted examples. There a certain volcanic neck filled with diamond-bearing breccia has been mined out to a depth of 3500 feet. This neck narrows somewhat in depth, that is, it is funnel shaped; but other diamond-bearing necks were found to grade downward into dikes and the breccia that fills the necks to change into massive rock. The interpretation put on these facts is that magma rose along a fissure and, having arrived near the Earth's surface, exploded its way through to the surface Presumably, such volcanoes were short lived. Longer-lived volcanoes are doubtless fed through conduits extending down to a magma reservoir.

The "Roots" of Volcanoes. Volcanism, as we have seen, is only one effect, the surface manifestation, of the rise of magma from deep within the Earth's crust. The present volcanism throughout the world is the continuation of an activity that has persisted since the beginning of geologic time. The present sites of activity are not necessarily those of earlier geologic time. However, it is remarkable how long volcanic activity has persisted more or less continuously in certain areas—certain parts of the crust have functioned as safety valves over millions of years.

Although magma has risen to the Earth's surface and has been ejected at volcanoes and has also been discharged in colossal amounts during mass eruptions, much of it failed to reach the surface. It halted within the rocky crust, where on solidifying it formed bodies of intrusive igneous rock, ranging from insignificant to enormous dimensions. Such intrusive masses are the laccoliths, stocks, and batholiths previously described. The conduit of every extinct volcano, could it be traced downward, would be found to join an intrusive mass below, or, if the conduit is feeding an active volcano, to extend into a body of magma—the volcanic reservoir—which in time will solidify as an intrusive mass.

Life and Distribution of Volcanoes

Age of Volcanoes. The span of life of active volcanoes differs greatly in different individuals. Etna, as we know from written testimony, has been erupting for the last 2500 years as it does now. Its great volume compared to its slow rate of growth makes it probable that at least 300,000 years have been required to build this grand volcano. From the human standpoint that is an immense lapse of time. On the other hand, we know from geologic evidence that Etna did not begin to erupt until the middle of the Pleistocene, the most recent of the geologic time epochs. From the geologic standpoint Etna is a youthful volcano.

It is impossible to judge whether a volcano is extinct or is merely dormant, because hundreds of years may elapse between eruptions. In the Middle Ages, Vesuvius had been dormant so long that its crater was overgrown with vegetation and it gave no sign of life. But in 1631 it became violently eruptive and has since been almost continuously active.

New Volcanoes. A few volcanoes, about ten, have come into existence during historic time. The first of these is Monte Nuovo near Naples, which was born in 1538. In a few days it built a cone of trachyte pumice 450 feet high, but discharged no lava. At the end of a week it ceased activity and has never resumed.

Izalco in Salvador, which began in 1770, blew out fragmental material at first, but soon ejected lava, which continued for five months. The volcano has been almost continuously active ever since and has built a cone 2600 feet high. In its short life Izalco has already discharged more lava than any other Central American volcano, and it ranks among the world's most active volcanoes.

Violent activity broke out in 1937 in Blanche Bay, the principal harbor of New Britain in the South Pacific. Two volcanoes, Matupi and Vulcan, were simultaneously in eruption. Vulcan was a newcomer and in a day or so built a cone 700 feet high. The volcanoes threatened to overwhelm Rabaul, the capital town; consequently the Government gave orders that the town be evacuated, and its 4500 people were moved to a safer place.

The infant among volcanoes is Paricutin in Mexico, 125 miles due west of Mexico City (Fig. 216). On February 19, 1943, some 300 earthquakes occurred. Next day the new volcano was born. It blew out in the field of Dionisio Pulido, who had gone there to pasture his sheep.

OTTO BREHME.

Fig. 216. Paricutin, the world's youngest volcano, in violent activity.

According to his account, about noon he saw the first smoke issue in thin white columns. By four o'clock flames burst out accompanied by clouds of dense smoke, explosions, and loud noises. After that, activity rapidly intensified. By the fifth day the lusty young volcano had already built itself a cone 300 feet high and was exploding continuously at the rate of 17 explosions a minute. Ten days later it was 722 feet high. It kept growing in height, and on its first birthday was 1410 feet high. Early in the life of the volcano basic lava began to flow out from near the base of the cone. Other flows followed. The temperature of these lavas near the source was 1200°C. Spectacular effects were common, especially at night, when there were magnificent displays of fireworks. During the early months the volcano blew out mainly incandescent bombs, generally 3 to 5 feet in diameter, occasionally 15 feet. Red and orange as seen at night, the bombs falling on the cone and rolling down its flanks illumined the whole volcano in a mass of fire.

Paricutin is in a region containing scores of large basaltic shield volcanoes, as well as thousands of basaltic cinder cones, all of about the same size as Paricutin. Jorullo, a new volcano that came into existence in 1759, breaking out in the middle of a plantation, is 45 miles southeast of Paricutin; after it had attained a height of 1300 feet it became inactive.

Paricutin, as well as all the other new volcanoes, has broken out in a volcanic area. The probability is therefore small that any volcano in the future will break out in a nonvolcanic area. However, four hundred years of immunity for the nonvolcanic areas, which is the span of time on which the prognosis is based, is insignificantly short in the long-range view of geology. In the light of the fact that volcanoes in the geologic past have broken through any part of the Earth's crust regardless of its constitution, the possibility exists that at some time volcanoes will break out in areas now nonvolcanic.

Volcanic Activity at Lassen Peak. No well-authenticated volcanic eruption was witnessed within the limits of the United States (outside of Alaska) until May, 1914, when Lassen Peak in northern California erupted explosively. The eruptions were chiefly of gases and rock fragments. In 1915 two enormous explosions took place—two great blasts that blew out horizontally from beneath the lava plug in the top of the volcano and laid waste a wide swath of country. Lassen Peak is a decadent volcano, whose explosiveness appears to be stimulated by water from melting snow gaining access to the still-liquid magma that is solidifying and crystallizing within the conduit. In recent years the only activity has been the feeble emission of gases.

Geographic Distribution of Volcanoes. The present active volcanoes exceed 500; but those dormant or but recently extinct, as shown by their slightly eroded condition, amount to several thousand.

Volcanoes tend to occur in long belts on the Earth's surface. The most marked of these belts encircles the Pacific Ocean. It extends northward along the Andes, through Central America into Mexico, through the United States and Canada to Alaska, then along the Aleutian chain to Asia, and, turning southward through Kamchatka, Japan, and the Philippines, it crosses the East Indies, and by various island chains again passes into the Pacific. Portions of this belt, like the Andes and the Aleutian chain, are remarkably linear and well developed (Fig. 217). Linear arrangement of volcanoes occurs not only on this large scale, but also on a small scale, as shown by the occurrence of isolated groups of volcanoes along lines.

Volcanoes occur both on the continents and in the oceans. They are notably abundant in the Pacific; some are extinct or dormant, but many are active; in fact, three-fifths of all active volcanoes are in the Pacific alone. Here also many of the volcanoes occur along lines and stand on submarine ridges that rise from the deep-sea floor. A superlative example is the Hawaiian ridge, extending for 1800 miles and crowned with mighty volcanoes, active and extinct. The linear arrangement of volcanoes suggests that they are as a rule situated on, or near, lines of weakness marked by fracturing or by folding in the Earth's crust (p. 400).

On the other hand, a volcano, or a group of volcanoes, may originate without being connected with fractures or with axes of folding. Volcanic forces have been sufficiently powerful to bore or blast their way to the Earth's surface without the aid of a fracture in the outer crust.

That almost all active volcanoes are situated either in the sea, or around its borders, and if inland are in or near lakes, has led many to believe that there must be a necessary connection between surface waters and the cause of volcanic activity. This question is considered later in the discussion of the origin of volcanoes.

Subaqueous Flows; Pillow Lavas. From the prevalence of volcanic islands in the sea, it is evident that vast outpourings of lava have occurred on the sea floor. The volcanic chain of the Hawaiian Islands is an example of this. Eruptions in progress beneath the sea have been recognized by the issuance of vapors and ash from the water. Thus, in 1831 a volcano was formed in the midst of the Mediterranean Sea, making a new island called Graham's Island. Being composed of loose material, it was soon reduced to a shoal by the waves. The three

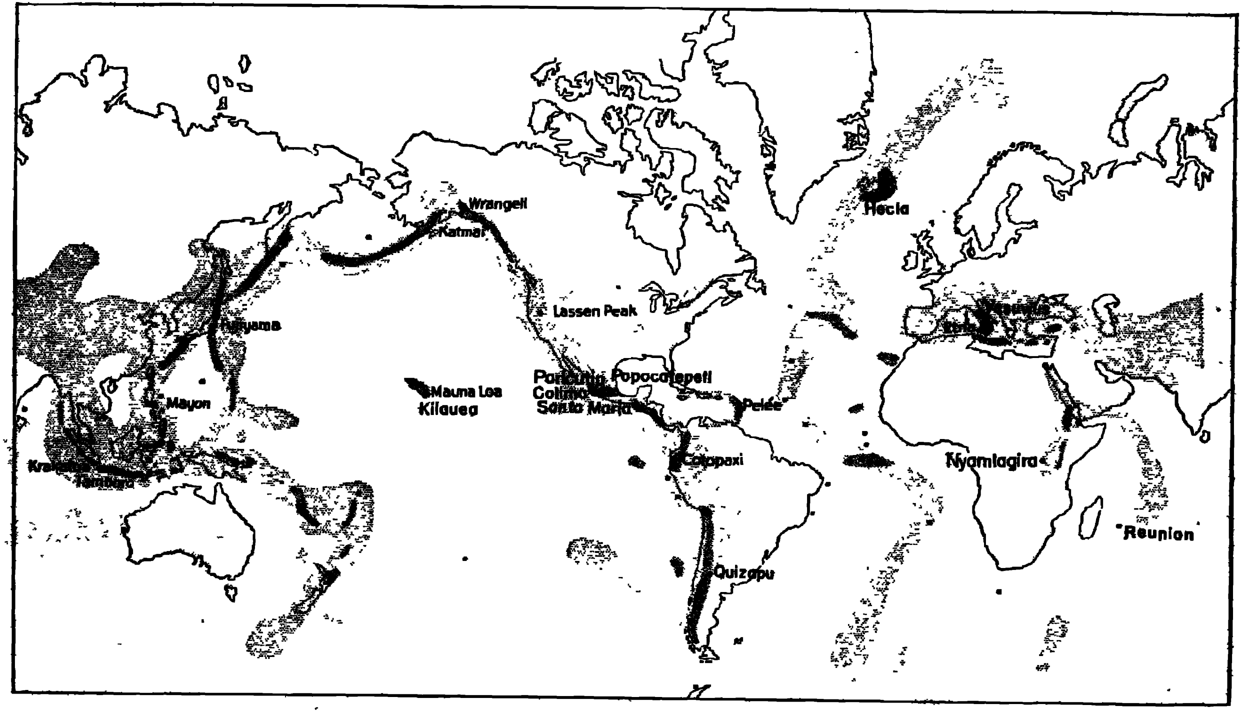


FIG. 217. Distribution of volcanoes and of earthquake belts throughout the world. Active and extinct volcanoes are shown by small circles. Areas subject to frequent earthquakes are shown by shading; zones of especially strong seismic activity are indicated by the darker shading.

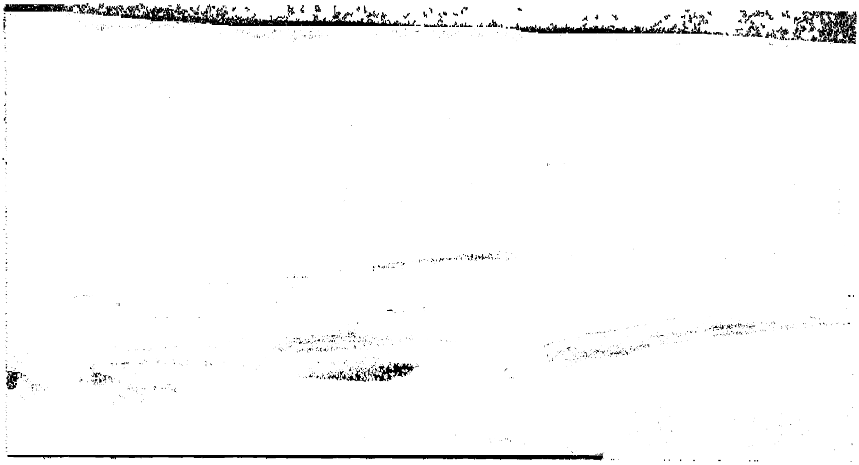
Bogoslov volcanoes in Bering Sea formed in 1796, 1883, and 1906 are other examples. In 1934 two new volcanoes appeared above the sea near Japan; one, in the north, is basaltic and grew to a height of 400 feet; the other, in the south, is rhyolitic and stands 80 feet above sea-level.

Eruptions occur also on the floors of bodies of fresh water. Moreover, streams of lava flow from the land into the sea or into lakes. Basaltic lavas that were erupted subaqueously or that flowed into bodies of fresh or marine water may take on a curious internal structure during solidification. The resulting lava flow resembles a pile of pillows, the individual pillows along their contacts showing mutual indentation and accommodation to the shapes of their neighbors (Fig. 218). Such lavas are expressively termed *pillow lavas*. Consequently, when we find pillow lavas intercalated between sedimentary rocks, we infer that they were erupted into a body of water.

Sidney Paige, U. S. Geological Survey.

FIG. 218. Pillow lava, the result of eruption under water.

Mass Eruptions. Lava has been poured out on a gigantic scale at intervals since the beginning of geologic time and has deluged enormous tracts of country. These floods of lava were discharged from deep-going fissures in the Earth's crust, that is, from fissures not connected with volcanoes, and were erupted with little accompanying explosive activity. By these outpourings immense broad plains or plateaus have been formed consisting of superposed horizontal sheets of lava with a few intercalated beds of tuff.



SIMMER STUDIO, WENATCHEE, WASHINGTON.

Fig 219. Basaltic lava, showing the horizontal superposition of flow upon flow. The cliff is 2000 feet high. Columbia River, near Trinidad, Washington

A basalt plateau formed in this way is the immense lava field of the Columbia Plateau in Washington and adjacent portions of Idaho and Oregon, which is 200,000 square miles in extent. As shown by the deep canyons of the Columbia River, Snake River, and other tributaries, the lava pile is as much as 4000 feet thick (Fig. 219). The Deccan trap (basalt) of western India now extends over a compact area of 250,000 square miles, but erosion has been at work ever since it was erupted 60 or 70 million years ago and has detached many areas as outliers. Originally the Deccan trap extended as a continuous basalt plateau over nearly 500,000 square miles. In places the piled-up lava sheets, which are individually thin, averaging about 15 feet, total 10,000 feet in thickness.

The horizontal layering of the lava sheets that make up these plateaus testifies to the extreme fluidity of the issuing magma, which permitted the lava to flow on a nearly level surface for many miles before congealing. Another result of the easily fluent nature of the lavas is that the average flow is only 15 to 40 feet thick. In the vast quantities of lava that have issued as mass eruptions to form the plateau basalts, in the Columbia Plateau and Deccan as well as many other regions, we see the most voluminous effects of volcanism.

Although basalt is the lava chiefly poured out from fissures by mass eruption, rhyolite has been poured out on a similar grand scale in certain regions.

ORIGIN OF VOLCANISM

Origin of the Heat. The downward increase of temperature, measurable in tunnels, mines, and boreholes, indicates that at depths greater than 25 to 50 miles the temperature will exceed that of known magmas, 1000° to 1200°C. According to a view long held, this heat is original, in the sense that it is residual from the formerly molten state of the globe, which in spite of its great age is still intensely hot at moderate depth.

The discovery of radioactivity has shown that some of the chemical elements are spontaneously disintegrating and that during this process they are giving off heat in notable quantity. Uranium and thorium are the elements of most importance. Consequently if there are local concentrations of radioactive matter in the crust, the heat generated by radioactivity in those places will gradually accumulate after the lapse of millions of years, melt the rock matter, and produce bodies of magma. According to this idea the radioactive changes that are going on within the Earth produce the heat necessary for volcanic action and even for the vaster igneous manifestations represented by the enormous batholiths that rise from the depths during epochs of mountain making.

Radioactive matter is universally present in all rocks, in extremely minute concentration. However, if the rocks deeper in the crust contain as much radioactive matter as those at the Earth's surface, the amount of heat evolved will be so large that the Earth will not be cooling off, as it was hitherto thought to be, but will actually be heating up and in time will become molten. It is certain, however, that the amount of radioactive matter diminishes greatly in depth. Manifestly, the radioactive content of the Earth's crust is of profound importance to its thermal condition, but it is too early yet to come to positive conclusions based on these revolutionary discoveries.

Origin of the Diversity of Magmas. Why the Earth has yielded, and is still yielding, so great a variety of magmas is a highly interesting problem. Different volcanoes erupt lavas of different kinds. Vesuvius erupts one kind, Etna another. More remarkable still, neighboring volcanoes may erupt very unlike lavas. Stromboli and Vulcano are in the Lipari Islands north of Sicily, only 25 miles apart: Stromboli is ejecting basalt, but Vulcano in the recent past has ejected rhyolite—lavas about as far apart in composition as possible.

Furthermore, a volcano during the course of its life may erupt lavas of several kinds. As many as five kinds of lava were erupted from San

Francisco Mountain, a large extinct volcano rising 5000 feet above the plateau of northern Arizona. The lavas that Vesuvius is now erupting differ greatly from the material ejected by its ancestor Somma. Evidently an evolution has been going on through the centuries in the reservoirs from which the volcanoes were fed. During a period of long repose of a volcano the composition of the magma in the reservoir may change, so that, when the volcano erupts again, lava of a different kind is ejected. What causes the changed composition of the magmas erupted? It is believed that originally a magma of uniform composition filled the reservoir beneath each volcano and that by a series of internal changes this magma altered in composition.

The process by which a magma of initially homogeneous composition changes is called *magmatic differentiation*. This process accounts not only for the diversity of lavas erupted, but also for the variety of igneous rocks that make up the intrusive masses—the laccoliths, stocks, and batholiths. The methods by which magmas differentiate are many but need not be explained here. One, however, is readily visualized. During periods of repose, the top of the lava column in the pipe of the volcano crusts over. Below this crust, the more "basic" minerals crystallize out slowly and, being heavier, sink in the still-liquid portion. Thus there accumulates a heavier, less silicic layer overlain by a lighter, more silicic layer. When the volcano erupts again, the first material ejected is more silicic, and the later material less silicic. An extreme illustration of this principle appears to be given by the 1931 flow of the active volcano on Reunion, an island in the Indian Ocean. In that year it discharged a long stream of rapidly flowing lava, more than half of whose bulk consisted of large crystals of olivine, although the normal basalts contain olivine not visible as a rule to the unaided eye. The great abundance of olivine in the lava of 1931 was the result of the settling out of these crystals toward the bottom of the reservoir during a period of quiet; later, the magma thus enriched in olivine was erupted.

Magmatic differentiation within the volcano's reservoir accounts for the variety of lavas erupted from one volcano. It does not account, however, for the fact that the reservoir of each volcano was initially charged with magma of differing composition. Differentiation within deeper-seated bodies of magma is invoked as an explanation. The composition of the magma with which the reservoir of the volcano is initially charged is determined by the stage of differentiation that had been reached in the deep-seated parent mass when some of it was expelled to a higher level in the crust to fill the volcanic reservoir.

The parent magma from which the diverse magmas were derived is thought to be basaltic in composition. One of the cogent reasons for this belief is that the great mass eruptions that have discharged such enormous volumes of lava at intervals throughout the whole span of geologic time are of basaltic composition. An ample source of basaltic magma in depth is thus demonstrated. Probably a layer of basalt in a potentially liquid state underlies everywhere the visible crust. We say "potentially liquid" because the evidence from the study of earthquakes, as shown in Chapter 16, suggests that the material at the depth of the supposed basaltic layer has rigidity, a property foreign to liquids. Because of the tremendous pressure in the Earth's depths, the material, although very hot, is probably not liquid but rigid. Therefore, if the pressure on the layer of potentially liquid basalt is relieved at any place, as for instance by upward buckling of the Earth's crust or by reduction of the superincumbent load as the result of deep erosion, or by both, melting would ensue, and a body of magma would form.

Origin of the Gases. The chief magmatic gas is superheated steam (referred to also as water vapor, or simply as water). It generally makes up 95 per cent or more of the total content of gases dissolved in the magma. This water may have been part of the original substance of the Earth, entrapped in the molten rock matter at or shortly after the time the Earth was formed. It may have been in part formed by the union of hydrogen, an original constituent of the magma, with oxygen of atmospheric or other origin; or it may be water that was absorbed by the magma from the surrounding rocks; or it may have been acquired by the magma in melting up rocks containing water-bearing minerals. A quantitative evaluation of these possibilities is beyond the present powers of science. A magma that has dissolved much limestone has become highly charged with carbon dioxide, which, after superheated steam, is the most abundant volcanic gas. Such absorption of limestone would generate enormous pressures, and such local development of pressure may have determined the sites of certain volcanic vents. Vesuvius, as already mentioned, is believed to owe its explosive power in part to this process.

The combustible gases—hydrogen, sulphur, hydrocarbons—appear to be original constituents of the magma. Their combustion, especially that of the hydrogen, produces the only true flames seen at an eruption, and the heat liberated makes the magma somewhat hotter than it is at greater depth in the vent.

That many volcanoes are in or near the sea or near lakes was formerly considered to be strong evidence that the gaseous water contained in the magma was obtained from descending surface water. But this evidence when examined loses part of its force. The nearness of some volcanoes to the sea, like those of North and South America, is only relative to the sizes of the continents. Actually they are far inland: in South America 100 to 250 miles, and this includes some cones still active—like Cotopaxi—which are not near any inland water body; in North America 30 to 130 miles or more. Although most of these are extinct, most of them when active were not near any bodies of water. However, volcanic cones are permeable edifices, and ground water can move freely through them; and some volcanologists regard it as highly probable that magma standing in the crater and supply pipe of a volcano absorbs water from the surrounding rocks and thus augments enormously its explosive potentiality.

Cause of the Rise of Magma to the Earth's Surface. Two analogies have long been used in explaining the rise of magma to the Earth's surface. The escape of lavas, said Dutton in 1880, "is analogous to what takes place when a bottle of warm champagne is suddenly uncorked." The energy of the imprisoned gases is thus considered to be the prime motive power. For the volcanoes that are fed from shallow reservoirs this still appears to be the most reasonable explanation. The other analogy is that of the cracked ice pond, with the water flooding out on the ice. According to this idea, the gross weight of the overlying cover of solid rocks forces the magma upward. The explanation thus suggested appears to fit eruptions of the type illustrated by the Hawaiian volcanoes, which are fed not from reservoirs within the crust but from the basaltic layer below the crust, and it appears to account particularly well for mass eruptions.

FUMAROLES AND HOT SPRINGS

During the dormant periods between eruptions many volcanoes give off steam and other gases. Long after a volcano has ceased to be active these emanations may continue to issue from its crater or from its flanks. Furthermore, large bodies of magma that rose to higher levels in the crust without breaking through to the surface have, while solidifying, given off their dissolved gases. These gases escape through fissures to the Earth's surface. The gases issue at the surface either as gases or, if sufficiently cooled, as liquids. The escaping gases will be considered under the general heading of *fumaroles*, the liquids under *hot springs*.

Fumaroles. *Fumaroles* (from the Latin *fumare,* to smoke) are fissures or holes in the rocks from which steam and other gases escape. The fume or "smoke" of the fumarole is therefore mainly steam, which generally forms 99 per cent of the total. Other gases, such as carbon dioxide, hydrochloric acid, hydrogen sulphide, hydrogen, methane, and others, also occur. Fumaroles that give off sulphurous vapors are termed *solfataras,* from the Italian word for sulphur.

Besides the substances already mentioned, the gases emitted by some fumaroles carry metals, such as iron, copper, and lead. These metals have been rendered volatile by the presence of chlorides and fluorides in the magma and consequently are able to leave the magma and escape into the surrounding rocks. As they approach the Earth's surface the volatile metallic compounds begin to react with the other fumarolic gases and are deposited as metallic minerals in the fissures through which the gases are streaming. Hematite is probably the commonest mineral formed in this way. During an eruption of Vesuvius a fissure 3 feet wide was thus filled with hematite in a few days. Galena, the chief ore mineral of lead, is occasionally formed at Vesuvius as the result of reaction between the lead chloride and the hydrogen sulphide emitted from the fumaroles. When this happens, ores, as it were, are actually being deposited under our eyes. In fact, the contemplation of these phenomena on the flanks of Vesuvius first suggested the fruitful idea that there is a genetic relation between igneous rocks and the occurrence and origin of ore deposits throughout the world.

The gases issuing from some fumaroles are exceedingly hot. In the remarkable fumarole field known as the Valley of Ten Thousand Smokes, which came into existence in 1912 at the time of the catastrophic eruption of the volcano Katmai in Alaska, temperatures as high as 645°C. have been measured. Some of the fumaroles occurring in the Valley of Ten Thousand Smokes are shown in Fig. 220.

Volcanoes that discharge only gases are said to be in the *fumarolic stage,* or, if the gases carry sulphur compounds, in the *solfataric stage.* Some of the great cones of the Cascade Range, such as Shasta, are in a feebly solfataric stage. Although the steam given off by a fumarole is probably in the main derived from the magma, its amount is likely to be increased by subsurface water that becomes vaporized, either by contact with hot rocks or by the hot gases of the fumarole itself.

Carbon dioxide gas is given off copiously in many places of active volcanism and in many places where volcanism has long ago died out. In some places carbon dioxide issues directly from the ground as a gas spring. Being heavier than air, the carbon dioxide collects during still

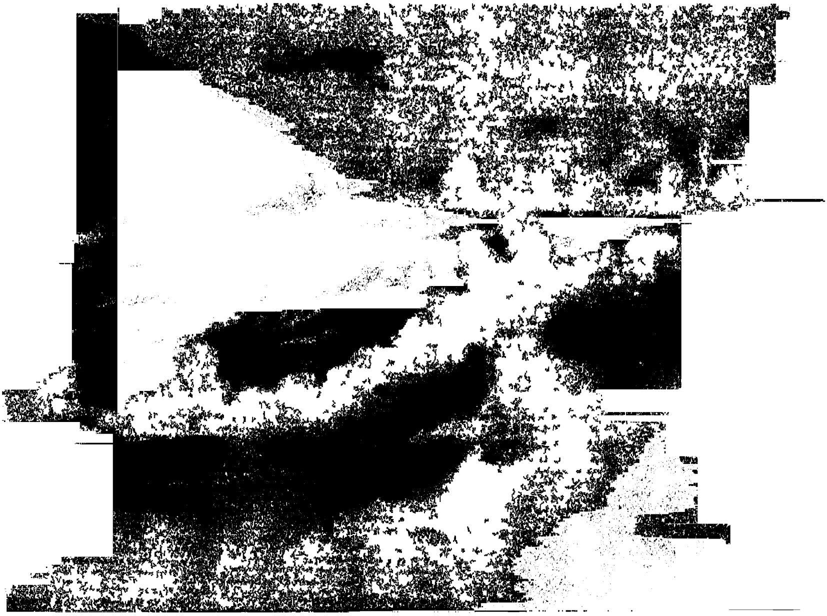


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Fig 220 Fumaroles issuing from the floor of the Valley of Ten Thousand Smokes, Alaska.

weather in depressions near the vent; and, as it is colorless, tasteless, and odorless, pools of this gas are veritable death traps for animals that happen to enter them Their deadliness is attested by "Death Gulch" in Yellowstone Park, where animals as large as grizzly bears have become asphyxiated.

Carbon dioxide given off by magma solidifying in depth may come in contact with ground water, which thereby becomes charged with the gas. If this carbonated water passes through limestone, it readily dissolves the limestone. When it emerges as a spring at the Earth's surface and the carbon dioxide escapes, it deposits the dissolved calcium carbonate as travertine. In this way immense masses of travertine have been formed by Mammoth Hot Springs in Yellowstone National Park Not all travertine deposits, however, are made by hot springs whose carbon dioxide is of magmatic origin. As shown on page 131, some travertine deposits have been made by springs whose carbon dioxide is of nonvolcanic origin.

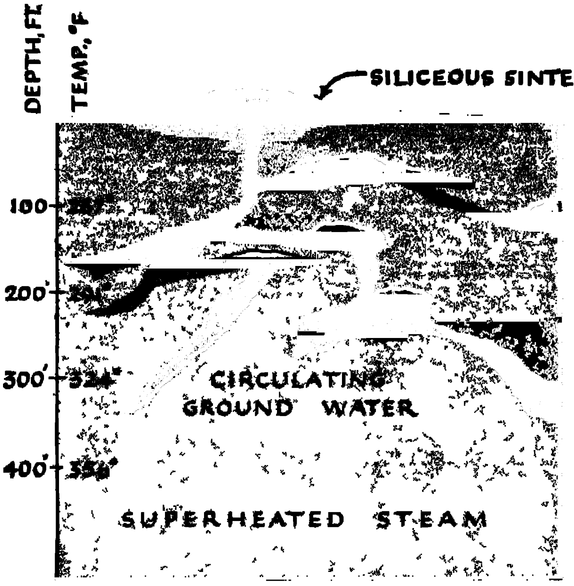
Hot Springs of Volcanic Regions. Hot springs as well as fumaroles abound in many volcanic regions. The thermal springs that have no connection with volcanism, such as those at Hot Springs, Virginia, have already been mentioned (p. 125). In many volcanic regions as the dry season comes on some of the hot springs become fumaroles, and when the wet season returns the fumaroles become hot springs. This evident seasonal alternation leads to the theory that hot springs are fed chiefly by ground water that has become heated by magmatic steam. If the descending ground water that enters the fumarole fissure is too abundant for the heat supplied by magmatic steam to vaporize it, the fumarole becomes a hot spring; in short, such a hot spring is a "drowned" fumarole. The circulatory system of a hot spring is in principle like the hot-water heating system in a house, but, instead of a furnace in the basement supplying the heat, magmatic steam furnishes the heat necessary to keep the hot spring flowing. The greater weight of the downward-moving column of cold water forces the warmer, lighter column of water to move upward to the exit at the surface, where it issues as a hot spring.

It is impossible in general to tell how much of the water (and steam) of hot springs and fumaroles is of surface origin and how much is of magmatic origin. The proportion contributed by each source doubtless differs in different regions. The hot springs in Lassen Volcanic National Park and Yellowstone National Park are estimated to consist of 10 per cent of magmatic water and 90 per cent of water of surface origin.

That magmatic steam has contributed to the hot springs of volcanic areas is indicated by the presence in the waters of such substances as arsenic, boric acid, and other constituents in quantities and under conditions that show that they could not have been dissolved out from the surrounding rocks of the country. The water of the hot springs of Yellowstone Park is largely of surface origin (about 90 per cent, as already mentioned), which becomes heated in depth by the condensation of magmatic steam in it and returns in this heated condition to the Earth's surface.

Although there are many kinds of hot springs, according to temperature and substances in solution, the most interesting are boiling springs and geysers.

Boiling Springs. Boiling springs are a feature of many volcanic regions. They are abundant in Lassen Volcanic National Park and in Yellowstone National Park, especially in the geyser basins. They grade from pools that are hot but rarely boil, or else simmer quietly, into springs that boil strongly and steadily. Some even boil violently and somewhat explosively, interrupted by short periods of repose. The violently boiling springs form transitions to geysers. As long as a spring has a sufficient supply of water to maintain an overflow, it remains limpid and is usually deep blue or green. But if the spring by its boiling evaporates the water as fast as it comes in, the water becomes more or less turbid from particles of disintegrated rock and eventually becomes a mass of boiling mud. Such hot springs are called "paint pots," or "mud pots."



Modified from E. T. Allen and A. L. Day.

FIG. 221. Vertical section to show the conditions necessary for geyser action. White areas represent large cavities in which steam accumulates.

Geysers. A geyser is a hot spring that at intervals erupts a column of hot water and steam. Geysers have marked individualities of their own, as is strikingly shown in Yellowstone National Park. Some geysers eject a column of water but a few feet high, such as the beautiful Sapphire Pool, which is a geyser almost suffocated but still attempting to erupt every 10 minutes. Others blow out columns as high as several hundred feet. The highest geyser in Yellowstone Park is the Giant, 250 feet. The world's record height was achieved by Waimangu, in New Zealand, which erupted to a height of possibly 1000 feet, but the very intensity of its eruptions appears to have ended the life of this geyser in the short span of 3 years. The eruptions of some geysers last a few seconds, in others they range up to several hours; and the water discharged ranges from a small quantity to hundreds of thousands of gallons. Some geysers erupt at regular intervals, but most erupt at irregular intervals, ranging from minutes to hours, days, weeks, or even longer: in some, years elapse between eruptions. Geysers are not common; in fact, almost all are in three regions: Yellowstone National Park, Iceland, and New Zealand.

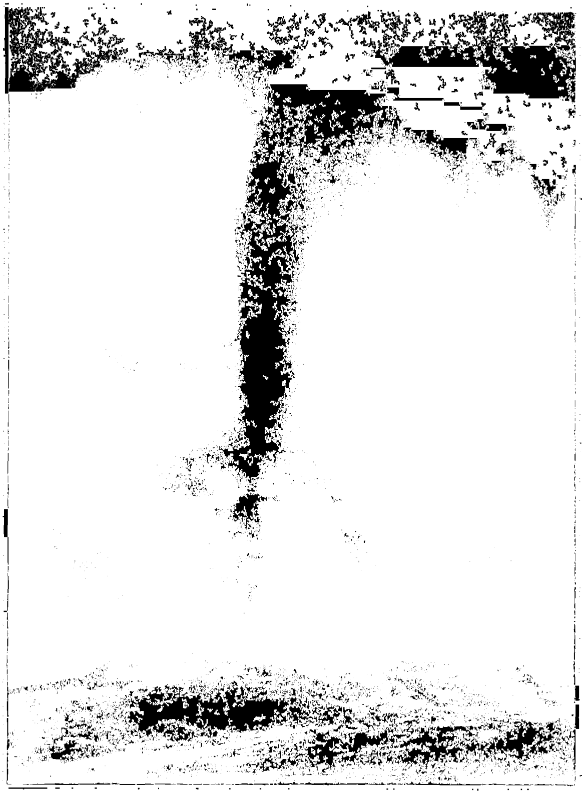
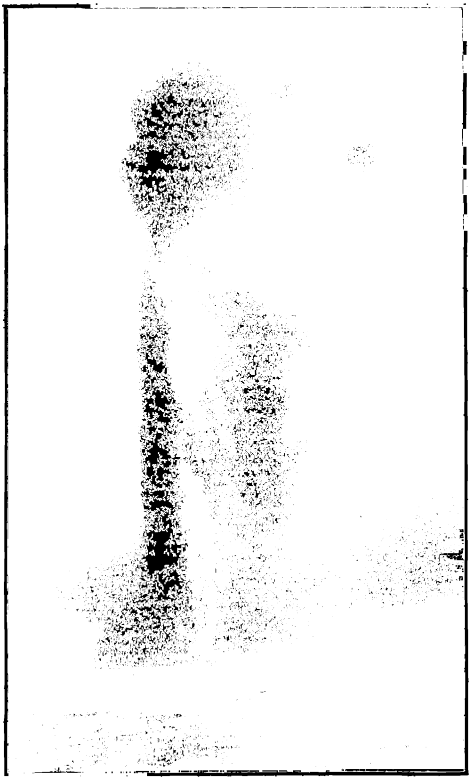

C. D. Walcott.

FIG. 222. Siliceous sinter cone of Lone Star Geyser. Yellowstone National Park.

Some geysers erupt into pools, which are several feet or yards across and may be rather deep. Most of these pools are ornamented around their sides and in the adjacent overflow areas by deposits of opaline silica, called *siliceous sinter* or *geyserite*. From the orifice of every geyser a tube extends downward in depth as shown in the diagram (Fig. 221). The tube that feeds a geyser pool is filled, except after each eruption, with water at or near the boiling point. Other geysers have no visible water in their tubes between eruptions and therefore have no pools around them. For all that can be seen at the surface they might be extinct. Some geysers have built cones of geyserite around their orifices, from a foot to several yards high, which form upward extensions of the tubes (Fig. 222).



Haynes Studio

FIG. 223 Old Faithful in eruption. Yellowstone National Park.

The most celebrated geyser in Yellowstone National Park is Old Faithful, so named because it erupts with almost clock-like regularity. It averages about 66 minutes between eruptions, very rarely going off schedule. It plays to a height of 110 to 160 feet (Fig. 223). About 10,000 to 12,000 gallons are blown out during each eruption. Year in, year out, regardless of season, winter or summer, it maintains this activity. A colossal amount of heat is thus brought to the surface by Old Faithful, not to mention that brought by the other 200 geysers in the Park. The only possible source of this heat is superheated steam given off by an intrusive mass of magma that is cooling and solidifying in depth.

The water of Old Faithful carries silica that it has dissolved out of the rhyolite through which it erupts, and this silica is being slowly deposited around the orifice as gelatinous silica, which hardens into siliceous sinter a few inches thick on the dome.

Since the Park first became known to the white man in 1871, a few geysers have become extinct or dormant, and several new ones have come into being. These changes do not mean that thermal activity in this region is likely to decrease in the immediate future, but merely that some changes in the underground system of pipes and fissures that supply the hot water are taking place. Altogether there are 200 geysers in the Park, and the hot springs, fumaroles, and other thermal vents exceed 3000. The amount of heat thus brought to the surface is enormous, being sufficient to melt about 3 tons of ice per second.

Cause of Geyser Action. The intermittent spouting of a geyser depends on the delicate adjustment of several factors: the amount and rate of inflow of subsurface water, the supply of heat, and the "plumbing," comprising the geyser tube and its underground connections of pipes and fissures. Because of so many variables, it is easy to understand why each geyser has its own idiosyncracies. As Allen and Day have pointed out from their long study of the geysers of Yellowstone National Park, the explanation of geysers is not so simple as generally considered.

The explanation of geyser eruption depends ultimately on the relation between pressure and the boiling point of water. The boiling point of water under the ordinary pressure of the atmosphere at sealevel is 212°F.; increase of pressure raises it, a decrease lowers it. Consequently, the boiling point at the bottom of a column of water is raised by the weight of the water above it. As shown in Fig. 221, the boiling temperature gradually increases in the tube of the geyser from the surface downward. If the tube is regular in form and large in diameter, convection currents become active as the water becomes hotter, thus mixing the water. Consequently it becomes nearly of the same temperature in all parts of the tube, and a boiling spring results. But if the tube is narrow or tortuous, the mixing of the water by convection is prevented, and the water reaches its boiling point in the deeper levels at the increased temperatures corresponding to the greater pressures. Suppose that at various points in Fig. 221 heat is being supplied by steam entering the pipes of the geyser. At first the steam condenses to water, thereby supplying a large amount of heat. Consequently the temperature of the water in the geyser system gradually increases until it is at the boiling point corresponding to the depth and pressure, when the steam can no longer condense but accumulates. Finally when enough steam has accumulated in this way, its expansive power lifts the water in the geyser tube until some overflows at the orifice, or, as in Old Faithful, until jets of water 10 to 25 feet high are blown into the air, thus heralding the coming eruption. These preliminary actions reduce the length of the column and thus lower the pressure on the water in depth; and this water, being now above the boiling point for the diminished pressure, flashes into steam, and a column of mingled steam and hot water is blown out of the tube into the air.

To account for the great quantity of water ejected by some geysers, much larger than the capacity of their tubes, it is necessary to assume that the tubes are connected laterally with one or more chambers, in which water and steam can accumulate.

That the geysers and other hot springs of the Park derive their heat from a deep-seated source is shown by their occurrence not only along the shores but also within Yellowstone Lake itself, an immense body of cold water beneath which the rocks must have cooled to a considerable depth. The deep-seated source is probably a large body of consolidating magma, a batholith, from which superheated steam issues along fissures; and, where this ascending steam meets the ground water, hot springs and their most spectacular manifestation, geysers, are formed.

Economic Utilization of Fumarole Fields

Fumarole fields have been developed in recent years so as to yield large supplies of steam for power generation. The fumarole field of Tuscany, north of Rome, was the first to be developed. In an area of 140 square miles a large number of fumaroles are giving off superheated steam. From 1818 onward these fumaroles have been utilized to recover the boric acid that they are bringing up in solution. In 1904 a dynamo was driven by using the fumarolic steam, and subsequently large-scale development of this resource has taken place. Four central power houses, capable of generating 25,000 horsepower, have been built, and power is transmitted to Florence, 60 miles away, Pisa, and other cities.

The fumarolic steam is 94 per cent pure steam and 6 per cent by weight of other gases, chiefly carbon dioxide, which is being recovered and used as a refrigerant. It is planned to recover also the small admixed quantities of sulphur dioxide, methane, and helium. The steam is thus being fully utilized, both thermally and chemically. Wells have been put down to a maximum depth of 600 feet, and the flow as well as the temperature of the steam increases as depth is gained. Notable is the fact that the fumarole field is in sedimentary rocks, the nearest volcano being 15 miles distant. The source of the steam is thought to be at a depth of 15,000 to 25,000 feet, probably a large body of granite magma.

At "The Geysers" in the Coast Ranges of California, 75 miles north of San Francisco, is an area of 35 acres containing a few feeble fumaroles and some small but very hot springs. The name "The Geysers" is a misnomer, however, as none of the hot springs is eruptive. Some of the steam vents, by suitable control of their outflow, have been converted into artificial geysers, which can be made to play at the convenience of the visitor. In 1921 the idea was conceived of drilling wells in this area to develop a flow of steam for power purposes. Eight wells have been put down, the deepest being 650 feet, and copious supplies of superheated steam have been developed. In fact, it is estimated that four of the wells will on the average deliver more than 1300 horsepower each. As in Tuscany, the deeper a well is drilled the greater the flow of steam and the higher its temperature.

The power possibilities of the fumarole fields in Java also are being investigated. One of the fields was bored in 1926. The most promising well, 220 feet deep, yields steam sufficient to generate 1200 horsepower. It is reported that power is generated at one-third the cost of hydroelectric power. Many fumarole fields in Java, as well as others in the near-by islands of Sumatra and the Celebes, remain to be tested. Volcanic energy is being utilized on a small scale at Kilauea, in Iceland, and in Japan.

VOLCANOLOGY AND HUMAN AFFAIRS

The human race is extraordinarily pertinacious in continuing to live around volcanoes and even high up on their slopes. Vesuvius, "the pride and terror of Naples," is thickly surrounded by towns and villages and is covered with gardens, groves, and vineyards that extend far up toward its summit. Etna is cultivated up to an altitude of 4000 feet, intensively below an elevation of 1500 feet—orange and lemon groves, vineyards, and oleanders, all in vivid contrast to the black lava flows. Monte Rosso, 700 feet high, the largest cinder cone on Etna, built in 1669 during the most disastrous eruption in the history of Etna, is now green with vineyards half way to its summit.

Java, distinguished for its large number of highly active volcanoes, is nevertheless the most densely populated area of its size in the world. Here the social, or societal, aspect of volcanology has been most fully recognized. The Netherlands Government has instituted a Volcanologic Service, which by prognosis and warning of pending outbreaks and by defensive measures reduces the dangers of volcanism. The volcano Kelut broke out in 1919 after a dormancy of eighteen years. During the interval a crater lake had formed, and this water now rushed down the valleys, killing 5500 people. Tunnels have since been driven into the crater to drain it, rendering repetition of the disaster impossible. Coming eruptions are often indicated by local, "volcanic" earthquakes and especially by increasing temperatures of fumaroles. Instruments for recording earthquakes have therefore been installed at critical localities, and systematic measurements are made of the temperatures of the fumaroles. For Merapi, a notably active volcano, the records show that when the fumarole temperatures reach 600°C. there is danger of an eruption.

The bombing of a lava flow on Mauna Loa by the Army Air Force has already been recounted. It is the hope of volcanologists in Hawaii and in Sicily that, if proper measures are taken, such as, for example, construction of embankments, lava flows that threaten productive land can be diverted into unproductive territory.

A brilliant example of diagnosis and prediction was given by the volcanologist Perret during the eruption of Mont Pelée in 1929–1932. After the eruption had continued five months, during which hundreds of peléan clouds had been projected, he was able to make the reassuring announcement that, although the eruptive activity was likely to continue a long time before it would cease, nevertheless it would thereafter be tranquil. Based as it was upon an intimate knowledge of the volcano, this bold prediction was fulfilled. The optimistic opinion is ventured by Perret that if adequate studies are made an authoritative forecast can be given for the course of any other eruption, at any place.

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CHAPTER 15

DEFORMATION OF THE EARTH'S CRUST

Although the strength of rocks and the stability of "terra firma" have become proverbial in popular fancy, the Earth's crust is weak in a geologic sense, and at one time or another probably every part of it has been deformed or displaced. Within historic times, and repeatedly in the present century, there have been abrupt, catastrophic shifts along fractures that penetrate bedrock to great depth. From historic records it is known also that gradual movements have occurred, with results that are perceptible only after years or even centuries have passed. Back of human history we find in the geologic record of all continents evidence of ancient movements on a gigantic scale, which repeatedly changed the geography of the globe. The bedrock has been fractured, bent, and mashed, and in some mountain zones the rocks have been so changed by successive deformations that their original nature can only be conjectured.

All movements of the crust, resulting in relative vertical or horizontal changes of position and in deformation of rocks, are comprehended under the general term *diastrophism* (from the Greek, meaning *thorough turning*).

MOVEMENTS IN HISTORIC AND LATE GEOLOGIC TIME

Earthquakes. Abrupt movements in the outer shell of the Earth generate strong vibrations, which are felt as earthquakes. Scarcely a day passes without records of shocks from some part of the world, reminding us that the Earth's crust is unstable. Earthquakes are of such significance to science and are so important in human history that special attention is devoted to them in a later chapter (Chap. 16).

Changes of Level. The most conspicuous known movements of the land are in the vertical direction; and to determine the extent and rate of change it is necessary to have a convenient horizontal surface of reference. The average level of the sea is the most logical surface for this purpose, as the shoreline at mean tide level is essentially horizontal throughout its whole extent. However, it is not true that the sea surface is undistorted and permanently fixed. Adjacent to high continental borders the water is attracted laterally and upward by the land mass, and the water surface is slightly farther from the center of the Earth in such localities than it is along low, flat coasts or adjacent to oceanic islands. Moreover, the sealevel has varied within recent geologic times, first through withdrawal of much water from the sea during the accumulation of continental ice sheets, and later through restoration of this water by wastage of the ice (p. 195). If the water now locked up in the Antarctic and Greenland ice sheets should be liberated by the sudden advent of warmer polar climates, many of the world's greatest cities and rich lowlands would be covered by the sea. Furthermore there are reasons for believing that the sea has increased in size and depth through geologic time by the constant addition of magmatic water; and it is probable that the deep-sea basins have changed appreciably in size many times through upward as well as downward bowing of their floors and by slow filling in of sediment, with consequent changes of sealevel. But such changes are gradual, and their effects in shifting shorelines are essentially uniform all over the Earth; whereas many movements of the land are relatively rapid, and all such movements vary in amount from one place to another.

Elevation. The most striking proofs of uplift of the land consist of the locally elevated position of features that we definitely associate with the sea or its edge. Thus in many parts of the world outcrops of rocks with attached shells or skeletons of dead marine organisms, such as barnacles and corals, are found high above sealevel. A classic example of local changes in land level is furnished by the temple of Serapis built by the Romans near the seashore west of Naples. The three columns left standing have been bored by marine mollusks to a height of about 20 feet above the temple floor, and the shells of the animals remain in some of the holes. Moreover, along the shore in the vicinity of the temple there are sedimentary deposits containing abundant shells of species now living in the bay; these deposits are exposed in bluffs ranging up to 20 feet in height. It is clear, therefore, that after the temple was built the ground beneath and around it was submerged, and later emerged by at least 20 feet. If this submergence and later emergence had resulted from a general rise followed by lowering of sealevel, the marine sediments near the temple should be in evidence along the entire coastline of Italy and of all other lands. Since the evidence actually is restricted to a small area, we conclude that there was a local subsidence of the ground in the vicinity of the temple, and later uplift of about equal amount.

U. S. ARMY AIR FORCES.

Fig 224. Emerged marine terraces, formed on coral-reef limestone. Cultivated fields occupy the highest of three distinct terraces Aguijan Island, Marianas Group, Southwest Pacific (Scale can be judged by trees and the graded road.)

Strong testimony is given also by the abnormal position of conspicuous features made by wave work along a coast In parts of California, Chile, Scotland, and numerous other coastal regions, emerged beaches, wave-cut benches, and wave-built terraces make nearly level platforms, some of them hundreds of feet above present sealevel, terminated inland by wave-cut cliffs (Fig 224) Since these features do not have world-wide distribution, with uniform heights above sealevel, they indicate uplifts that affected limited segments of the Earth's crust

Other evidence, of a direct and positive character, is supplied by careful observations made year by year. Thus in the countries bordering the Baltic Sea an uplift still in progress has been under observation for a long period and has been measured by marks placed on the shores. In some places the elevation has been more than 3 feet in a century, but the rate varies systematically from one part of the region to another. Furthermore, raised beaches and other shoreline features indicate that this region has been rising gradually for a long period, so that parts of Sweden and Finland are at least 900 feet higher than they were at the close of the glacial ages (Fig. 225). Elevated shorelines are a noticeable feature in many northern regions. There is similar proof on a large scale that within recent geologic time the west coast of South America has experienced very considerable elevation; and probably the uplift is still in progress. Numerous destructive earthquakes within historic time testify to the instability of this coastal belt. A major disaster occurred January 25, 1939, when Chillan, a populous city in Chile, was almost totally destroyed.

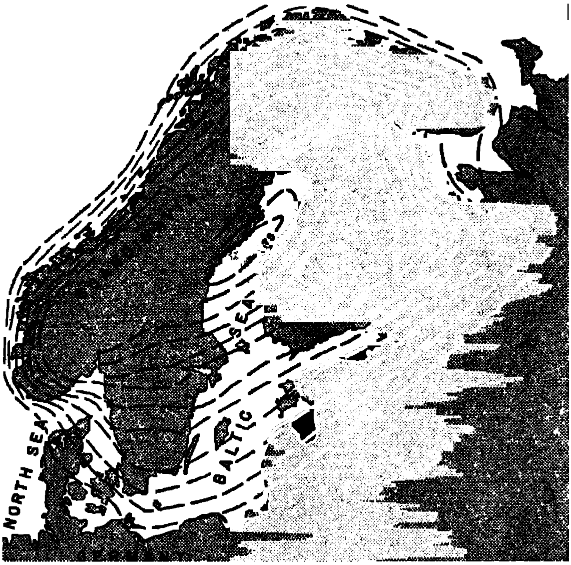

FIG. 225. Showing the domal form of uplift in Scandinavia and Finland after the disappearance of the last ice sheet. Shading indicates present land areas. The number on each of the heavy lines signifies the amount of uplift, in meters, experienced by every point on that line, as recorded by emerged shorelines and related features. Note that from zero elevation along the outer line the amount increases steadily inward to a maximum of 275 meters (about 900 feet) in northeastern Sweden. No lines are drawn in the central part of Scandinavia, because no emerged shorelines have been found in that mountainous area, which probably was not submerged.

Subsidence. Evidence of subsidence of the land below sealevel is less striking than that of elevation, but not less convincing. However, encroachment of the sea upon the land is not in itself a proof of subsidence, as it may result merely from cutting back of the land by waves and currents, or from a general rise of sealevel. Submergence of features that are definitely characteristic of land surfaces constitutes the best proof, if it is clear that the submergence is not worldwide but is limited in extent.

Evidence of submergence is furnished by some irregular shorelines formed by the drowning of valleys, with formation of bays and estuaries (p. 234). An excellent example is the coast of Maine, where the sea extends long distances up all the large stream valleys, and numerous hills on the former land surface have been isolated to form islands off the present coast. However, in many parts of the coastal belt of Maine patches of marine silt and clay, containing shells of species still living, are found more than 100 feet above present sealevel. Therefore there has been emergence also of that region, though not enough to overcome the effects of earlier submergence. Since the evidence of emergence disappears farther south, it reflects regional uplift and not a lowering of sealevel. This recent uplift of Maine and southeastern Canada was part of the widespread elevation that occurred after the weight of the continental ice sheet was removed from northern North America (p. 17).

In many parts of the world thick deposits of sediment are being laid down by streams in subsiding basins adjacent to coasts. Borings into some deltas pass through alternating marine and freshwater deposits, or even through sediments entirely nonmarine, to a great depth below present sealevel. For example, wells sunk into the delta of the Po near Venice pass through four separate layers that contain abundant remains of plants similar to those now growing in the marshlands along the Adriatic. One of the layers is about 300 feet below sealevel. At shallower depths some of the sands, gravels, and clays contain shells of marine mollusks, and other layers yield shells of land snails. In the delta of the Ganges, near Calcutta, pieces of wood and bones of land animals are found hundreds of feet below sealevel. Deep wells in the Great Valley of California furnish similar evidence. Facts of this kind suggest that subsidence has been going on for a long period, not at a uniform rate, but as an interrupted process whose variations permitted alternating freshwater and marine deposits to be formed.

Evidences of Elevation or Depression Inland. Movements of the crust involving changes of level are not confined to the sea coasts; they occur also in the interiors of the continents. For example, in 1811 large areas of the Mississippi floodplain near New Madrid, Missouri, sank far below their former level and now are occupied by lakes

(p. 142). Trees that grew on the plain were killed by the flooding, and for a long time their dead trunks or tops, projecting above the water, testified to the recency of the catastrophe.

An excellent illustration of tilting on a large scale is afforded by the Great Lakes. To the northeast the land has risen since the disappearance of the great ice sheet (p. 195), and as a result the lake basins have been tilted southwestward. Emerged shorelines are several hundred feet above the present lake surface on the north and northeast, and slope toward it as they are followed west and south. Since the lakes discharge to the east, the raising of their outlets has caused them to enlarge, expanding them to the west and south. The mouths of valleys on the south and west sides of some of the lakes (especially Erie and Superior) have been submerged (Fig. 226). The tilting movement is still in progress and has been accurately determined; it is at the rate of 5 inches per hundred miles per century. Small as this rate seems, in 1600 years it would cause the upper Great Lakes to discharge by way of the Chicago River into the Mississippi drainage.

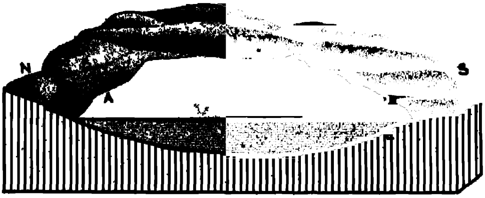

FIG. 226. Tilting of a lake basin from north (*N*) to south (*S*). *AD*, present lake level; *BC*, raised shoreline, disappearing under water at *C*; *CD*, evidence of submergence (drowned valleys); *E*, profile of former shore, now under water.

Horizontal Changes of Position. In addition to the vertical movements discussed above, important recent horizontal shifts have been detected. In 1906 the ground on opposite sides of a great fracture in California moved in opposite directions; the shift was indicated by abrupt offsetting of roads, fences, and similar features that crossed the fracture (Fig. 265, p. 396). Knowing that movements of this kind occur repeatedly in the Coast Ranges, the U. S. Coast and Geodetic Survey has established stations at critical points whose relative positions have been determined by precise measurements. At intervals of several years the measurements are repeated, and thus the directions and amounts of horizontal movements may eventually be determined.

Significance of Recent Vertical and Horizontal Movements. It is obvious that changes of position on the Earth's surface must be accompanied by bending or fracturing of the rocks of the crust. In fact, surface movements merely record deep-seated processes that lead to slight changes in the form of the globe. Commonly the movement is a broad flexing that affects wide areas; locally there is sharper bending, or the rocks yield by actually breaking. Probably some of the striking recent elevations, such as those observed in various lands bordering the Pacific Ocean, represent small steps in the uplifting of mountain ranges whose growth will continue for long ages.

In considering effects of this kind we are impelled to inquire at once into their ultimate cause; but discussion of this question will be postponed until more of the essential facts have been presented. It does not seem strange, however, that a globe so large as the Earth, which spins rapidly on its axis and whirls through space, which experiences chemical and physical changes in its outer and probably also in its deeper parts, should be subject to slow and almost continuous deformation. Volcanoes testify to local unstable conditions, which result in the melting and moving of rock material. Erosion and sedimentation through long periods result in the transfer of great loads at the surface, which undoubtedly sets up enormous stresses in the crust. These well-known processes would in themselves cause some slow deformation of the crust; and probably more profound changes in progress in the deep unknown parts of the Earth are responsible for still greater deforming forces.

ANCIENT CRUSTAL MOVEMENTS

It is neither desirable nor possible to distinguish sharply between recent and older deformation. Certain features at the surface, such as the marine terraces along the coast of Maine (p. 355), bear unmistakable witness to very recent crustal disturbances, even if there is no direct record of the events in human history. On the other hand, we see in the rocks many fractures and folds that date from very early periods in Earth history, as is shown clearly by their geologic relationships. Between these two extremes are found indications of disturbance in every geologic epoch, showing that movements of the same or similar kinds have been continuous or recurrent throughout recorded geologic time. In general, the latest movements are recorded in forms and features on the Earth's surface, such as the elevated beaches, submerged valleys, and tilted lake basins described above. All surface forms are ephemeral because of erosion; and, as they become fragmentary or disappear, the most reliable guide to former crustal movements is found in the structure of the underlying rocks.

EVIDENCE IN SURFACE FORMS

Certain topographic features that result from gentle warping of the crust develop slowly and persist for a very long time. For example, when a peneplane is arched up widely the rejuvenated streams become deeply incised, and their former meander patterns are thus preserved in the deepened valleys (Fig. 322, p. 491). As the deepest incision occurs in the area of greatest uplift, the middle portion and the edges of such a great arch can be recognized by study of the incised valleys. Furthermore, remnants of the uplifted peneplane persist for a long time between the valleys. By study of the valleys and the remnants of old erosion surfaces it is determined that the Appalachian region is an irregularly arched peneplane undergoing dissection. Evidence of this kind, used for recognizing warping movements that occurred in the geologic past, is similar to that by which recent uplift is determined.

Further explanation of surface features as indicators of crustal movement is given in the discussion of land forms (Chap. 20).

EVIDENCE IN SEDIMENTARY ROCKS

Thick masses of unconsolidated sediments, like those in the Great Valley of California, have been mentioned as evidence of crustal movement (p. 355). Similar evidence for earlier geologic periods is furnished by old sedimentary formations, especially by the thick sections exposed in dissected mountains or plateaus. In the Appalachian region the sedimentary strata, now greatly disturbed and eroded, are made in large part of sediments that were deposited on shallow sea floors near land. Although the total thickness of the strata is several miles, the evidence is conclusive that the sea in which they were deposited was never deep and that at times it actually disappeared, only to return again. Obviously there was continuous or recurrent subsidence of the area while the sediments were accumulating (Fig. 297, p. 463).

In some sections of sedimentary rocks a pronounced change in the character of sediments indicates either uplift or subsidence. Thus in the high plateaus of Utah a thick formation made of sand deposited by streams and by the wind shows that the region was above sealevel for a long period. Above the sandstone a formation of shale and limestone containing marine fossils records a period of widespread submergence. In the Catskill region of eastern New York a series of marine limestones and shales is succeeded by an enormous thickness of coarser-grained sediments, deposited chiefly above sealevel. Evidently the land adjacent to the old seaway was elevated, with the result that quantities of coarse debris were rushed into the basin where fine mud and limy sediment had been slowly accumulating.

"Breaks" in sedimentation, caused by uplift, erosion, and resubmergence, are of great importance in reading the history of a region. This aspect of sedimentary rocks is discussed later in the chapter, under the heading *unconformity.*

EVIDENCE IN STRUCTURE OF ROCKS

The most lasting effect of crustal movement is the disturbance of rocks beneath the surface. Sedimentary rocks are particularly useful in preserving the records of diastrophism because they are formed in nearly horizontal strata, and therefore even a slight bending or breaking is easily detected. Igneous rocks are in general less favorable for this purpose, since they are characteristically massive and irregular in their original form. Lava flows and ash beds are an exception, as they have some degree of layering, although this original structure is usually less regular than in sedimentary beds. Thin sills of igneous rock intrusive into sedimentary rocks bend or break with the strata and so help record the amount of deformation. On the other hand, the great size and the nearly uniform structure of granite batholiths make them poor recorders of Earth movements.

The principal types of structural features acquired by the bedrock from crustal movements are (1) broad warps, (2) folds, (3) fractures, (4) foliation. Since foliation results from fundamental change or *metamorphism* of rocks, it is explained in connection with that subject (Chap. 17). Careful attention to the results of deformation has great practical importance, since such a study is essential in locating and tracing valuable coal beds, mineral veins, and strata containing oil and gas. The structural features of the Earth have also a broader interest because they furnish clues to important historical events.

Broad Warps

A wide wooden board exposed to the weather becomes gently flexed or *warped* from its original flat form. On a much larger scale, rock strata are permanently warped by irregular uplift or depression of a region. In the Colorado Plateau of Utah and Arizona a thick blanket of old marine strata lies thousands of feet above sealevel and is dissected by deep canyons. In a general way these strata are nearly horizontal; but if any one layer is followed in detail it is found to bow upward into irregular domes and bend downward into shallow basins. It is clear that this wide area of sedimentary beds, many hundreds of miles across, was not lifted up with absolute uniformity, but was distorted somewhat. The original surface of the uplifted mass has been entirely destroyed by erosion; but a record of the distortion is preserved in the form of each layer of stratified rock beneath the former surface.

Warping is the commonest form of rock deformation near the Earth's surface. All the old marine strata that now lie on the continents are in some degree warped, where they have escaped more violent disturbance, because uplift of the former sea floors was greater in some places than in others.

Folds

Anticline and Syncline. In many places the stratified rocks have been buckled into more or less regular plications or *folds.* Some of these are on a small scale and can be seen directly; but commonly the folding is on such a great scale, and exposures of the rocks are so discontinuous, that it is necessary to study and piece together the structure of certain distinguishable layers over many miles before the form of the folds becomes clear. The nature and scale of the folds in mountain regions can be appreciated best by study of maps and cross-sections that have been prepared by geologists in the field.

Two terms are used constantly in descriptions and discussions of folds. Like regular swells on the sea, rock folds ordinarily occur in a series, with alternating crests and troughs.[1] The crests of the folds—that is, the upfolds—are *anticlines;* the troughs, or downfolds, are *synclines* (Fig. 227). Initially the anticlines form ridges, the synclines form valleys; but if the original surface crests are subsequently carried away by erosion, and the whole reduced to a nearly level surface, we still call the upfolded portions below the surface anticlines, the downfolded portions synclines, and in imagination reconstruct the missing parts. Thus it should be clear that anticlines and synclines *are features not of surface form, but of structure.* Commonly the original configuration of the surface is actually reversed by erosion, so that valleys now occupy the positions of the former crests, and ridges or mountains are in the places of the troughs; but the original structural terms still apply (Fig. 228).

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[1] Folding is distinguished from warping by this common pairing of crests and troughs, by greater regularity of plan, and generally by more pronounced bending of the strata involved.

It is cause for some wonder that strong, brittle rocks can be bent into sharp folds. If we should try to bend a rock slab by compressing the opposite edges in a powerful vise, the slab would be broken in two or crushed to bits. Probably two conditions are essential for making the rock folds found in nature. The tremendous forces that cause deformation act through very long periods of time, and the rocks yield slowly instead of fracturing abruptly; furthermore, the folded strata that we now see at the surface lay at some depth in the crust when they were deformed, and the confining pressure of the rock above and around them prevented the layers from breaking as the folding took place. Under sufficiently strong confining pressure the most brittle rocks can be forced to bend as if they were soft and plastic.

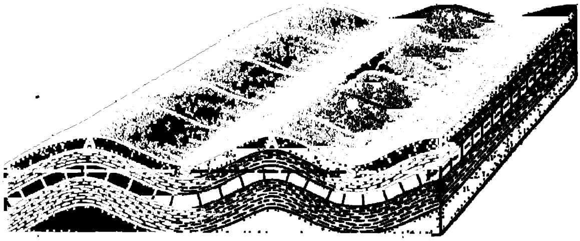


FIG. 227. Anticlines (*A*) and synclines (*S*). The diagram assumes that the folding occurred so recently that running water has had time to cut only a few small valleys in the anticlines; hence the anticlines form ridges and the synclines valleys. Even if erosion proceeds until a peneplane is produced, as represented by the broken line through *P*, the anticlines and synclines will still exist as structural features beneath the surface.

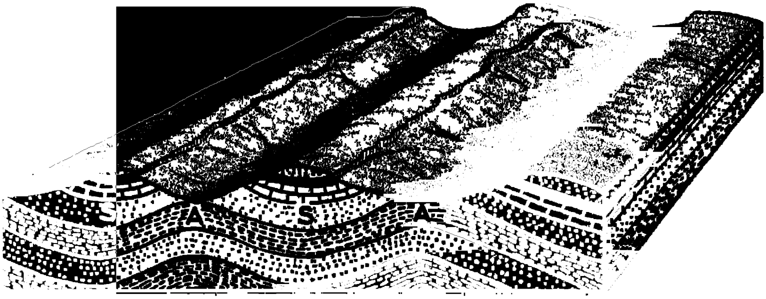


FIG. 228. Illustrating an advanced stage in erosion of folded strata. Anticlines (*AA*) have been cut down until valleys occupy their crests; a resistant stratum (black) protects the synclines (*SS*), so that ridges exist along these structural troughs. The original position of the resistant stratum across the valleys is indicated by the broken line.

Besides the bending of individual layers an important necessary adjustment in the folding of thick sections of strata is brought about by slipping of each layer over those adjacent. True folding is restricted to stratified rocks because in massive bodies, such as granite batholiths, there are no parallel planes on which adjustment by slipping can occur.

Method of Studying and Mapping Inclined Strata. Only the ideal relation of simple, upright, regular folds has been considered above. A series of folds approximating this form is by no means uncommon in nature, but usually the folding is much more complicated. The varied kinds of deformation which the rocks have undergone in any region determine the *geologic structure* of that region; and it is important, economically as well as scientifically, that the geologic structure of every country should be known as far as possible and represented accurately on maps. If the surface of the Earth were everywhere naked bedrock, this would be a comparatively easy matter; but, since the rocks have been greatly eroded and are largely covered with mantle and vegetation, or with water, snow, and glacier ice, the natural difficulties of the task are large. The structure in a region is determined by a careful study and comparison of the outcrops. If the ground were perfectly level and the strata horizontal, the outcrop would be the flat surface of the uppermost rock stratum, and we should learn little from it; but on slopes and cliffs bordering stream valleys we may inspect the outcropping edges of many horizontal strata, as in the Grand Canyon (Fig. 11; and Fig. 290, p. 454).

On the other hand, if the strata have been inclined by folding and later eroded, their edges are exposed even on a nearly flat land surface. Commonly the edges of the harder, more resistant beds project to form the more prominent outcrops In mountain districts, particularly in arid regions, soil and other concealing debris usually decrease in amount with increasing height; and exposures of rock grow in prominence correspondingly, until each of the upper rocky ridges and peaks is a vast outcrop (Fig. 116, p. 172). Because of the excellent exposures and the great depth of the section visible in canyon walls and on cliffed slopes, mountains furnish the most favorable opportunities for determining geologic structure.

Strike and Dip. The inclination of a rock layer is called the *dip*. If the stratum is resistant, its projecting edge trends across flat country as a definite ridge (Fig. 229). The direction of this ridge is called the *strike* of the stratum. In precise terms, *the strike is the direction of* *the line formed by intersection of the plane of bedding with the horizontal plane.* In Fig. 229 the strike of all the strata and of the dike *D* is north. The dike and the two strata *A* and *B* are resistant and so form prominent ridges whose directions are easily determined.

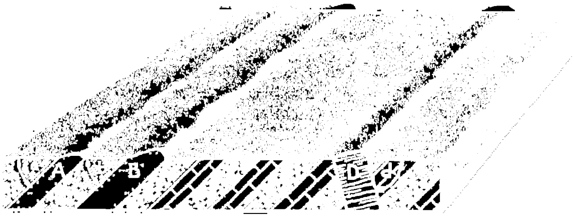


FIG. 229 The *strike* of two resistant strata, *A* and *B*, and of a dike, *D*, is indicated clearly by the trend of the ridge formed by each. The strike in each case is exactly north. *A* and *B* dip west (the dip angles are *a* and *b*); the dike dips east (angle of dip *d*).

The inclination or *dip* of a rock layer is the angle between the plane of bedding and the horizontal plane. In Fig. 229 the surface of the ground is supposed to be horizontal except for the three projecting ridges. Then the angle *a* (50°) gives the amount of dip of layer *A*, *b* (50°) of the layer *B*, and *d* (80°) of the dike *D*. But it is essential to give the direction as well as the amount of dip. *A* and *B* dip 50° *west*, *D* dips 80° *east*. The direction of the dip is at right angles to the strike.

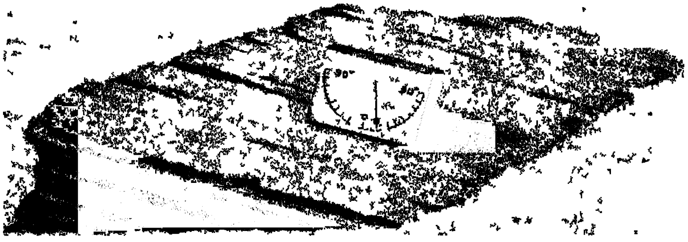


FIG. 230. Measurement of dip angle with a clinometer. The pendulum of the instrument swings freely on an axis, and therefore is always vertical when the box is on edge. When the edge of the box rests on a bedding plane in the direction of dip, the dip angle (in this case about 16°) is read directly on the graduated arc.

The elements of strike and dip are better understood if they are compared with the gabled roof of a house. In such a roof the ridge-

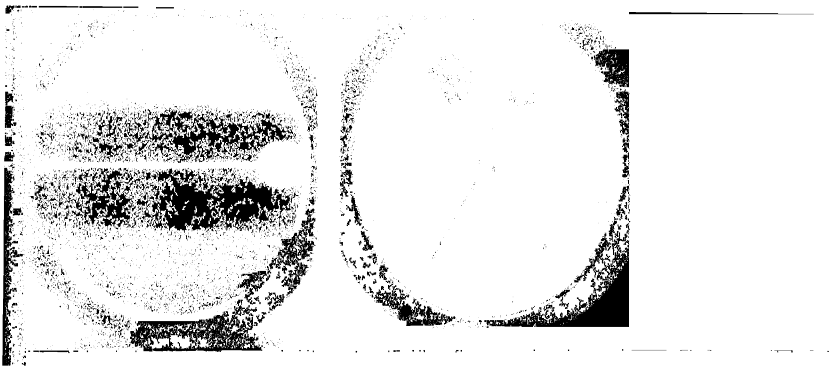


Fig. 231. Brunton compass with the lid open and sights extended. Mirror inside lid, with its dividing line in prolongation of the sights, is useful in reading directions. The compass circle is divided into quadrants, each graduated 0° to 90°. Semicircular arc on bottom of compass box is used in measuring angles of dip; when edge of instrument is on bedding plane in direction of dip, and level bubble is centered, reading on arc is the angle of dip. In the figure the reading is zero, since edge of instrument is horizontal. (⅔ actual size.)

pole, at the intersection of the two slopes, is horizontal. Let us call this line of the ridgepole the *line of strike*. If the opposite gables face exactly north and south, the direction of the strike is north. Thus the two halves of the roof have the same direction of strike. If each half slopes 45°, the *amount* of dip is the same for the two halves. But one half dips east, the other west.

The amount of dip is determined with a clinometer, which in a simple model is essentially a pendulum swinging over a graduated arc (Fig. 230). For geologic purposes the compass and clinometer are combined in one instrument, to permit ready determination of the strike and the amount of dip (Fig. 231).

The full procedure in determining strike and dip is illustrated in Fig. 232. At one edge of the inclined stratum *FGH* the compass shows the angle *NOH* between the edge of the stratum and the north-south line. If this angle is 45°, the strike is recorded as north 45° east (abbreviated N 45° E). The direction of dip is at right angles to the strike, or south 45° east (S 45° E), shown in the diagram by the angle *SOd*. With the clinometer the angle of dip, a, is found to be 60°, and the dip is recorded as 60° S 45° E. The meaning of this shorthand record is clear if it is kept in mind that the first angle—60°—is the *amount* of dip, measured downward from the horizontal, whereas the second angle—45°—gives the *direction* in which the stratum dips, in relation to the north-south line.

Dip and strike are represented on geologic maps by a conventional sign ⊢→, in which the direction of the bar, as placed on the map, indicates the direction of strike, and the arrow points in the direction of dip. Ordinarily the amount of dip in degrees is written in: for example, ⊢→ 30°. On many maps, the dip symbol is a plain line without the arrow point.

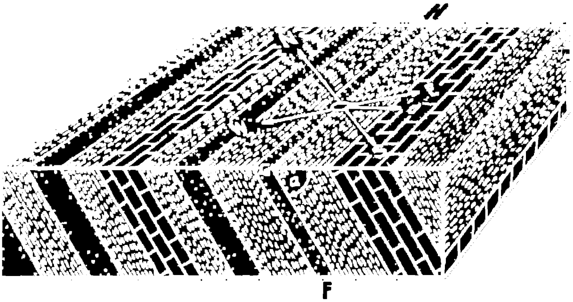


FIG. 232 Determination of strike and dip. The compass (*O*) is placed on the edge of a bedding plane of the stratum *FGH*. The arrows *N*, *E*, *S*, *W* indicate compass directions, *d* is the direction of dip, and the angle *a* is the amount of dip.

Elements of Folds. The sides of a fold are called the *limbs* (Fig. 233), and the median line between the limbs, along the apex of an anticline or the trough of a syncline, is the *axis* (Figs. 233, 234). This line extends along a bedding surface, or along this surface restored if it has been partly eroded. When folds that have horizontal axes are eroded deeply, the ridges made by the edges of resistant strata are nearly parallel as shown in Fig. 228; but if the axes are not horizontal, on a nearly flat erosion surface the outcropping edges of strata in the

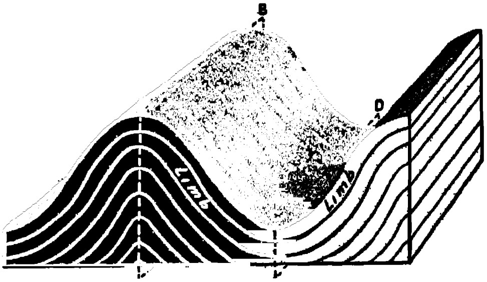


FIG. 233. Simple upright folds. The line *AB* is the axis of the anticline; *CD* is the axis of the syncline. The imaginary planes indicated by broken lines are the *axial planes* of the two folds.

limbs of any anticline or syncline converge and finally meet (Figs. 234, 235). The result is a hairpin turn in the outcrop in either kind of fold; but there are two general ways of distinguishing one from the other. In the ordinary syncline, dips are consistently inward toward the axis, and the younger strata lie inside the hairpin; in an ordinary anticline, dips are outward from the axis, and the older strata lie inside the hairpin (Fig. 234).

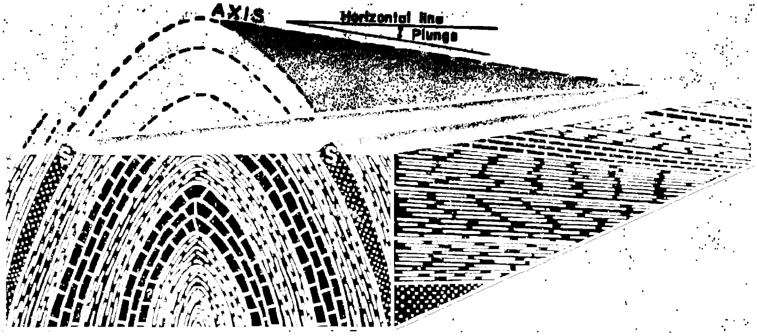
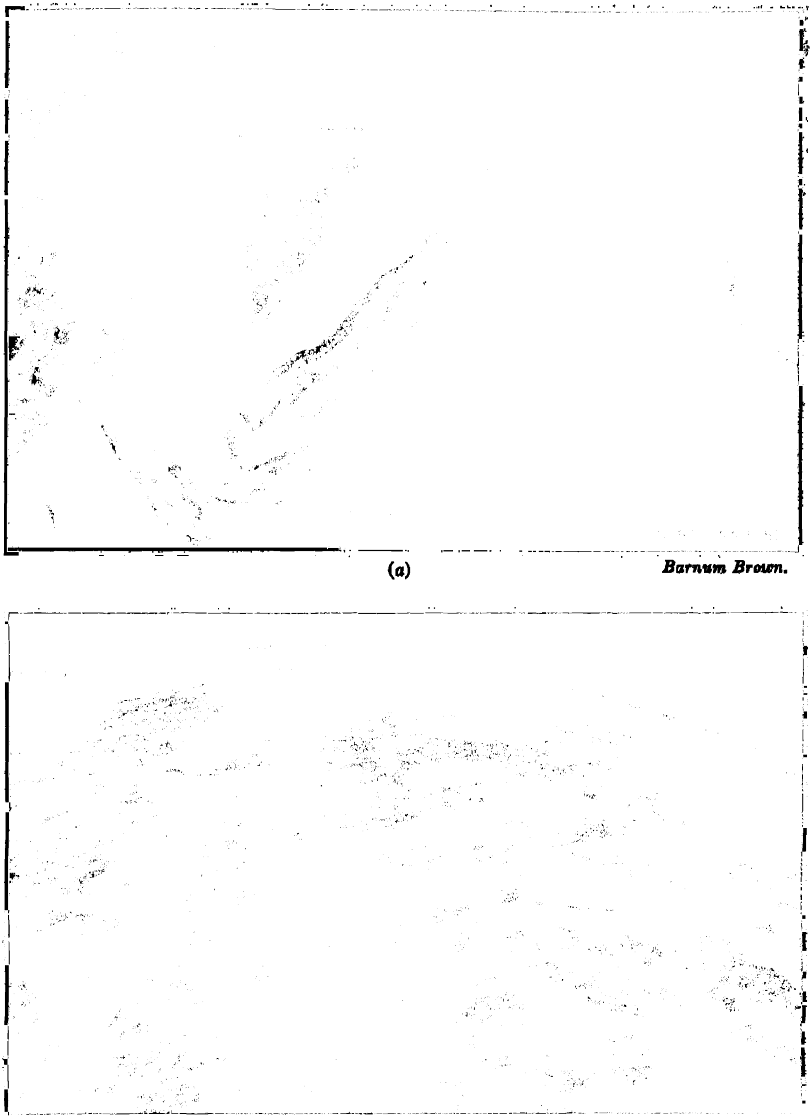


FIG. 234. An anticline tilted along its axis and eroded to a nearly level surface, except for a ridge marking the edge of a resistant sandstone layer. The surface of this layer as it appeared before erosion is shown by shading. The angle between the axis of the fold and the horizontal is the plunge.

Folds whose axes are not horizontal are known as *plunging* folds. The angle between the axis and the horizontal plane is called the *plunge* of a fold (Fig. 234).

Other Variations in Fold Structure. Thus far we have considered only regular, upright folds. An imaginary plane passed through the center of a fold and its axis, as in Fig. 233, is called the *axial plane* of the fold. In a regular or *symmetric* fold, this plane is one of symmetry; that is, the parts to left and right of it are symmetrically disposed, or each point on the left of the plane is the mirror image of a corresponding point on the right of it. If the fold is upright the plane is vertical (Fig. 234). However, some folds are not upright but have been pushed over until the axial planes are inclined (Fig. 7, p. 20; *6* and 7 in Fig. 236). In some folds the overturning has gone so far that the axial plane is nearly or actually horizontal; the fold is then termed *recumbent*. Great recumbent folds are among the complex features that characterize the Swiss Alps (p. 468). Such folds are found only in mountain belts that have experienced extreme deformation.


(a) *Barnum Brown.*

(b) *M. S. Kennedy.*

FIG. 235. Air views of large plunging anticlines.

(a) Sheep Mountain anticline in the Big Horn Basin, Wyoming. The older strata are resistant and form a high ridge at the core of the fold; higher strata have been eroded from the top of the anticline, and the more resistant of them form the low ridges on the flanks. (Compare Fig. 234.)

(b) North Dome, Kettleman Hills, California, looking northwest. Resistant strata make ridges, which outline the flanks and the plunging end of the great fold.

Some overturned folds break at the apex or on one limb; and on breaking, the parts are likely to be displaced with respect to one another, or *faulted*. Faulting, however, is so important a phenomenon that it is given special consideration in later pages.

If folds are so sharply flexed that the limbs dip at least 45°, they are said to be *closed;* in this condition the horizontal distance across the strata, or the width of the fold, can not be further reduced without squeezing or mashing of the beds (*3* to *7* in Fig. 236). If the limbs make a large angle with each other (as in *1* and *2*, Fig. 236) the fold is *open*, and the strata can be further folded without mashing.

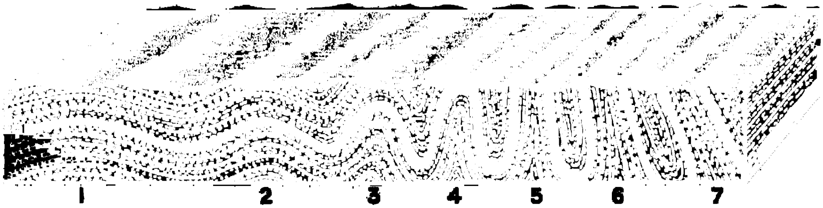


FIG. 236. Folded strata, considerably eroded. The anticlines *1* and *2* are *open* folds; farther to the right the folds are *closed*. At the right of the block the folds are *isoclinal*. The layer *BBB*, which can be traced through the entire series of folds, serves as a useful key; broken lines indicate parts of this layer removed by erosion. Length of block about 25 miles.

In *isoclinal* (equal-inclination) folds the strata are compressed until, on both sides of a fold, and perhaps throughout a series, they are parallel and have the same dip (Fig. 236, *5* to *7*). When the crests of such folds are cut away by erosion, considerable skill is required for correct interpretation of the structure, since there is no difference in general appearance between anticlines and synclines. The general principle followed is to determine the positions of axial planes by study of the strata. The oldest exposed strata are repeated at the middle of an eroded anticline, and the youngest strata at the middle of a syncline. For example, in Fig. 236 the layer *B*, seen at the left, can be traced entirely through the series of folds. In the anticline *6*, layers below and hence older than *B* are repeated on opposite sides of the axial plane; in the synclines adjacent to *6*, layers above and hence younger than *B* are similarly repeated.

A special kind of fold is the *monocline*, in which the strata are bent in one direction only (Fig. 237). A true monocline is a one-limb flexure, on either side of which the strata are horizontal or have uniform gentle dips. Monoclines are particularly well developed, as isolated features, in the nearly flat strata of the Colorado Plateau in northern Arizona and southern Utah.

Economic Aspect of Folds. Correct solution of problems encountered in folded strata is of the utmost importance to mining geologists and others concerned with the recovery of mineral products. In many oil fields the occurrence of petroleum and natural gas has a definite relation to flexures in sedimentary beds, as explained in a later chapter (p. 517). In Fig. 236, the folded black layer *BB* may be taken to represent an important coal bed in the Appalachian region. From hasty surface examination, it may appear that there are six or more separate coal beds, each extending downward indefinitely. A skilled geologist, however, will recognize the evidence of folding and will be able to estimate the quantity of available coal. In the famous Homestake gold mine of the Black Hills, South Dakota, the formation containing the ore is complexly folded, and precise determination of the structure is necessary for profitable mining as well as for any dependable estimate of the quantity of ore still in the ground. Innumerable other examples illustrating the practical importance of folds could be cited from every country in which mining and well drilling are practiced.

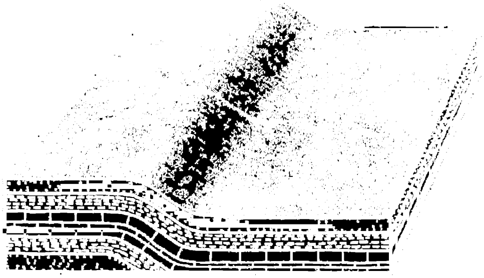


FIG. 237. A monocline. It is assumed that the fold developed rapidly and recently; therefore it is little eroded. Length of block about 2 miles.

Geosynclines and Geanticlines. Long belts within a continent or on the sea floor have been warped down to form *geosynclines*, which range in width from scores to hundreds of miles. Correspondingly great upwarps are called *geanticlines*. The prefix in each case (from the Greek word *geos*, meaning Earth) emphasizes the scale of these features. In contrast with ordinary folds, the flexures responsible for geosynclines and geanticlines are very gentle.

The geosynclines of the past have been the great basins for the accumulation of sediments like those exposed in the Appalachians (Figs. 297–300, p. 463) and the Alps. Slow downwarping of the floor of a geosynclinal basin continues as sediments accumulate, until in the middle portion the thickness of deposits reaches several miles. It is only by the later uplift and dissection of these deposits that the original nature of the geosyncline is revealed. Thus a geosyncline has peculiar features other than its great size to distinguish it from an ordinary syncline in a belt of folding. A geosyncline is not formed by simple downbending of sedimentary strata already formed; an essential element is continued deposition of sediments on the sinking floor of the basin. Thus the first layers of sediment to be deposited become gently flexed downward as the movement continues; but strata formed during the later stages may be nearly horizontal (Fig. 297).

Geanticlines also grow slowly, and their crests are eroded away as the upwarping continues. A good example, on a moderate scale, is the Cincinnati Arch in Ohio and Kentucky, where the strata dip gently away from a median axis in an area 250 miles wide.

Joints and Faults

In the outer shell of the Earth the rocks are traversed in all directions by fractures, which vary from microscopic crevices to important breaks on which large displacement has occurred. Such fractures have great geologic value, because they are clues to important events in the history of a region, and particularly because of the aid they give to various geologic processes. We have considered the large influence of fractures on the weathering and erosion of rocks (Chaps. 3, 5) and on the circulation of ground water (Chap. 7); and we shall consider their importance also in connection with mineral deposits (Chap. 21).

A fracture on which there has been no appreciable displacement parallel with the walls is a *joint*. If there has been relative displacement of the walls parallel with the fracture, so that corresponding points on the two sides are distinctly offset, the fracture is a *fault*. Many joints and faults are closed so tightly that little or no space is visible between the walls. If the walls are distinctly separated the term *fissure* is sometimes applied to the fracture. Some fissures are open, and others have been filled with mineral matter deposited by circulating solutions.

Joints in Sedimentary Rocks. Field examination shows that joints are common, but that they are much more numerous in some places than in others. Where they are abundant, commonly they are arranged in more or less definite *sets;* in each set the joints are nearly parallel. In many places there are two prominent sets of joints, intersecting at a large angle. Such a combination of two or more intersecting sets constitutes a joint *system.* Combined with natural bedding planes, a well-defined system of joints divides stratified rocks into series of closely fitting blocks. The finer the grain of the rock, as a rule, the more perfect the jointing and the more sharply defined the resulting blocks. In some beds of shale and limestone the jointing is exceptionally perfect (Fig. 238).

Some jointing probably results from the tension caused by contraction in beds of sediments when they are elevated from the sea floor to form land masses and lose part of the contained water as a result. A more common probable cause of regular fracturing is the warping and twisting experienced by the strata during crustal movements In regions where the strata have been folded, as in many mountain zones, probably the force that folded the beds also produced many joints. Some sets of joints extend for long distances across a thick series of

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Fig 238 Joints cutting nearly horizontal beds of limestone There are two sets of joints, nearly vertical and at right angles to each other (Exposed planes of one set are entirely in shadow; in the other set they are largely in sunlight) Drummond Island, Michigan

beds and are known as *master joints.* They are contrasted with minor fractures, some of which are limited to a single stratum.

Joints in Igneous Rocks. A common type of jointing in igneous rocks is caused by the contraction resulting from the cooling of the igneous material during and shortly after solidification from the liquid state. It may manifest itself in one of several ways, depending on the rate of cooling, the size and shape of the igneous body, and other factors. Large intrusive masses of granite and similar rocks are characteristically cut by joint planes that divide them into large blocks or prisms. Finer-grained rock in sills, laccoliths, and dikes commonly is divided into small angular fragments by closely spaced joints. Some laccoliths and similar dome-shaped intrusions have a shelly jointing on a large scale, parallel to the domed surface. This appears to have been caused by nearly uniform cooling of the mass from the periphery, with resulting separation into sheets.

The most striking kind of contraction jointing in an igneous rock produces columnar structure, which is characteristic of tabular masses. Many dikes, sills, and lava flows are made up of closely fitting prisms subdivided by less conspicuous cross-joints. The prisms have a variable number of sides, but commonly they tend to be hexagonal, particularly in lava flows, and some of them have remarkable regularity of form. They range from several inches to a number of feet in diameter, and up to 500 feet in length. The Giant's Causeway on the north coast of Ireland is one of the most celebrated examples of this columnar structure. The columns form at right angles to the chief cooling surfaces, and consequently in a level intruded sheet or a flow of lava they are essentially vertical (Fig. 239), whereas in a vertical dike they are nearly horizontal. Thus some dikes, exposed as walls by erosion, resemble regularly piled cordwood. In igneous bodies that have curved surfaces the position and form of the columns depend on the form of the periphery of each individual mass.

In addition to the joints caused by cooling, later fractures caused by crustal movements also affect igneous rocks; but wherever a prominent columnar structure or other well-defined fracture system is original in the rock mass, later stresses are more likely to be relieved along these existing breaks than to form additional fractures.

Jointing in Metamorphic Rocks. As a rule, metamorphic rocks are much jointed. This might be expected because of the extensive deformation to which such rocks have been subjected. The character of the jointing varies considerably with the nature of the rock. Many of the massive gneisses have joint systems like those characteristic of

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Fig. 239. Columnar structure in a thick lava flow. The Devil's Post-Pile, Middle Fork of the San Joaquin River, California.

granite, whereas rocks that retain original bedding, such as quartzite, have joints similar to those found in sedimentary rocks.

Practical Importance of Joints. Joints are a matter of great importance in all quarrying, tunneling, and mining operations where rock work enters as an important factor, since the jointing obviously facilitates the work of excavation. Without them, every rock fragment would have to be broken or blasted loose from bedrock. However, joints may also be a serious inconvenience, especially if large blocks of quarried stone are desired. Perfect monoliths 50 to 100 feet in length can be obtained from comparatively few localities.

General Features of Faults. Displacement of rock masses along a fault (p. 370) may occur at the time of the break, or at some later time. Faults are common features in rocks of all kinds. They are most evident in stratified formations, since the offsetting of definite layers makes a break conspicuous and directly measurable. However, massive igneous rocks are faulted as well; and, as mineral veins and other features of economic value are displaced by such fractures, it is extremely important from a practical as well as a scientific standpoint that the nature of faults be well understood.

A surface of fracture along which movement and dislocation have occurred is often spoken of as a fault plane. Although a limited part of it may be nearly plane, it is rarely flat for any considerable distance, but more or less curved and irregular. Therefore it is better, and causes less misapprehension, to term it a *fault surface*. Rather commonly a faulting movement occurs, not upon one surface, but upon a number of more or less closely adjacent breaks, producing a *fault zone*, in which the various offsets make in the aggregate the total displacement. The masses of rock involved in fault movements generally are of such size and weight, and are so compressed together, that the motion of one fault face on the other takes place under tremendous pressure. As a result of the friction the rock faces are smoothed and striated, and not uncommonly receive a high polish. Such polished and grooved surfaces are known as *slickensides* (Fig. 240). The direction of the line formed by the intersection of the fault with the plane of the horizon is called the *strike* of the fault, just as we speak of the strike of inclined strata. The surface of faulting is rarely exactly vertical; usually it is inclined, and in some important faults it ap-

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Fig. 240. Part of an old fault surface, with slickensides, uncovered by erosion. The striae and flutings indicate that the movement was directly down the dip of the surface. Spotted Range, Nevada.

proaches horizontality. The angle between the fault surface and the horizontal plane is the *dip*. In an inclined fault the side that overhangs is known as the *hanging wall*, the other as the *foot wall* (Fig. 241). If one were to descend along a fault, as in an inclined shaft of a mine, the appropriateness of these old mining terms would be evident.

Generally the fracture is closed tightly; but parts of it may have been open at one time and have been filled with mineral matter deposited from solution. Along many faults the grinding of the walls upon one another has produced a zone of broken and crushed rock known as *fault breccia*. Commonly the finely powdered rock directly along the fault forms a seam of clay-like material, known as *gouge*. In the displacement of stratified rocks the friction usually causes bending of the layers near the fault surface. This feature, referred to as *drag*, is a useful aid in determining the relative direction of motion on the two sides of the fault (Fig. 241). Practically it is of the highest importance as an aid in finding the dislocated segment of a coal bed or a mineral deposit.

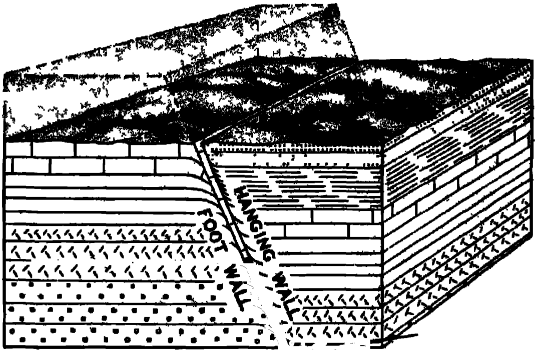


FIG. 241. Fault relations and terms. The strata have been displaced by a fault, and a vein of mineral (black) has formed along the fracture. Mining operations have removed the mineral to a considerable depth, exposing the hanging wall and foot wall of the fault. Part of the upthrown side has been removed by erosion, as shown by broken lines. The strata were dragged upward in the hanging wall and downward in the foot wall by friction on the fault surface.

The features explained so far have to do chiefly with faults as seen below the surface of the ground. Ordinarily a fault breaks the surface as well as the rocks beneath; and, if one side of the fault is elevated with relation to the other, the result is a cliff, or *fault scarp* (Figs. 243, 254). With the passage of time the original scarp is modified or even removed by erosion (p. 384).

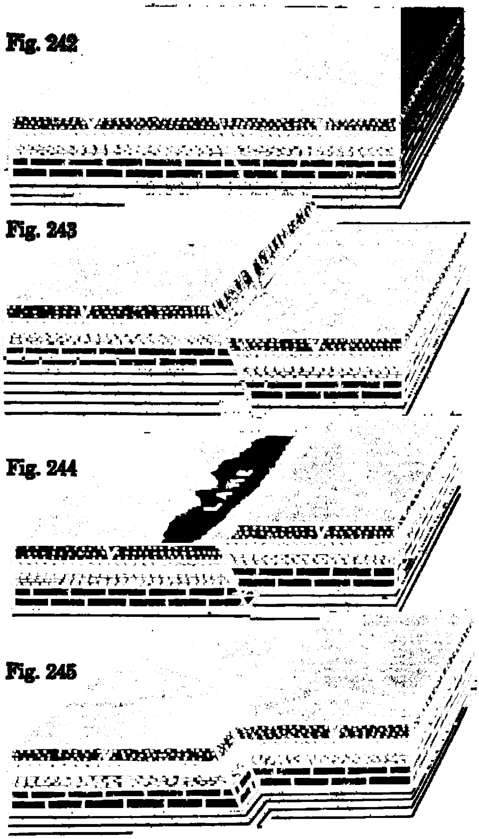
Movement on a Fault. Generally there is no means of telling how much actual movement there has been on either side of a fault. Even if a definite, recognizable object, such as a crystal or pebble in the rock, has been cut in two by the fracture and the halves carried apart a measurable distance, we can not determine whether one side of the fault stood still while the other side experienced all the displacement, or whether both sides shared in the movement. Conceivably we might know the exact position of every point on the ground before the faulting and later check the positions on each side of the break. Such a check is being attempted along some active faults in California. Generally, however, we lack the precise information necessary for such a study of modern displacements; and most faults with which we have to deal are old features whose original expression at the Earth's surface was long ago destroyed by erosion. The actual walls of such faults usually are seen only in mine tunnels or in other excavations of limited size.

Nevertheless an attempt is made to classify faults according to the apparent or relative displacement on the two sides as indicated by offset strata, dikes, or other recognizable features. Thus in Fig. 243, it is evident, from the displacement of the layer *V–V*, that the hanging wall has gone down with relation to the foot wall (although actually the foot wall may have moved up while the hanging-wall block stood still, or both blocks may have moved). A fault that has this relation between hanging wall and foot wall is called a *normal* fault. If the hanging wall appears to have been crowded up over the foot wall, as in Fig. 244, the fault is called a *reverse* fault.

Generally the movement on a fault surface is more or less oblique, as suggested in Fig. 246, instead of straight up or down the dip of the surface. Not uncommonly the motion is chiefly or entirely horizontal, parallel with the strike of the fault (Fig. 245). Such a fault is called a *strike-slip* fault. If it cuts only horizontal strata there is no measure of the movement except in displaced features on the ground, since the strata are not offset. Displacement of a steeply inclined dike, however, would record both the direction and the amount of movement.

The simplest possible examples are shown in Figs. 242–245. Strata and other features of bedrock commonly are inclined and hence are offset by either vertical or horizontal movement along most faults. Therefore the offsetting of rock layers, taken by itself, is not a safe criterion in solving fault problems. Some of the possible complications are discussed below and shown in Fig. 248.

Since generally we do not see much of the fault surface, it is necessary to calculate the displacement from the effects seen on the ground, on the side of a valley cut through the fault, or some other chance plane.



FIGS. 242–245. Types of faults. Relative movement indicated by the broken bed V–V.

FIG. 242. Before movement. Position of the fracture shown by broken line.
FIG. 243. Simple normal fault, making a scarp which the stream descends in a cascade.
FIG. 244. Reverse fault. The projecting edge of the hanging wall breaks off and slumps under its own weight. The stream is obstructed and forms a lake.
FIG. 245. Strike-slip fault, with no vertical displacement.

Therefore the displacement on a fault is described geometrically, by its components in three dimensions. These three components are named and explained in Fig. 246.

Faults in Stratified Rocks. Certain terms are used to define faults in relation to the structure of sedimentary beds that are affected. Thus in a *strike fault,* the strike of the fault and that of the strata are parallel, or nearly so (Fig. 247); *dip faults* cut directly across the strike of the strata, or nearly so (Fig. 248); *oblique faults* cut diagonally across the strike of the strata. The figures show only normal faults; but similar principles apply to reverse faults. The figures also

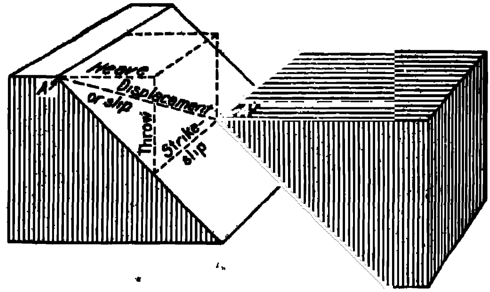


FIG. 246. Oblique movement on a fault; points A and A' have moved apart along the fault surface. Broken lines show the *displacement* (*slip*), and its three components—*throw*, *heave*, and *strike-slip*—measured along axes at right angles to each other.

indicate no real strike slip, but in Fig. 248, *C*, there is offsetting of beds, with a false suggestion of strike movement. Such abrupt offsetting of tilted strata is one of the strongest suggestions of dip or oblique faulting. Strike faults are more difficult to perceive and are easily overlooked; not uncommonly they cause deception as to the thickness of strata by producing repetitions (Fig. 247, *C*). On the other hand,

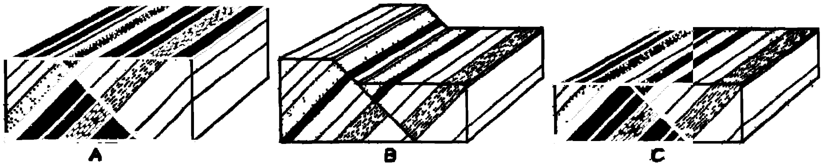


FIG. 247. Repetition of strata by normal strike faulting and later erosion. *A*, before faulting; *B*, directly after faulting; *C*, surface leveled by erosion.

some strike faults conceal strata after erosion has occurred. Such concealed strata may contain petroleum, or some valuable deposit of ore, which can be discovered only by a fortunate accident or by modern geophysical prospecting (p. 521).

Displacement on some faults is attended by rotary or pivotal motion. A fault of this general nature is known as a *rotary* fault; or, if the displacement dies out gradually up to a definite point, it is a *hinge* fault. After erosion has leveled the surface, a fault of this kind is indicated by a pronounced difference in the strike and dip of strata on opposite sides of the break. Some hinge faults pass gradually into monoclinal folds (Fig. 249).

Magnitude of Faulting. The scale of faulting varies within wide limits; displacement varies from a fraction of an inch up to thousands of feet, or even several miles. In the Plateau region of Arizona and Utah, several faults of great magnitude extend in a north-south direction, some of them crossing the Grand Canyon. Each of the largest fractures in this group can be followed 100 miles or more, and each has a throw measured in thousands of feet. The Great Basin region presents the phenomenon of faulting on a colossal scale. In the area between the Sierra Nevada on the west and the Wasatch Range on the east, the crust is divided into huge blocks by gigantic fractures; and differential displacement of these blocks, together with erosion, has resulted in mountainous topography (Fig. 291, p. 455).

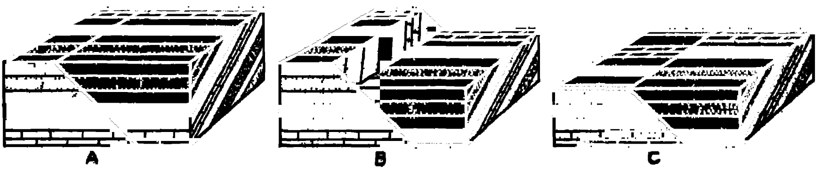


FIG. 248. Offsetting of strata by dip faulting and later erosion. *A*, before faulting; *B*, directly after faulting; *C*, surface leveled by erosion.

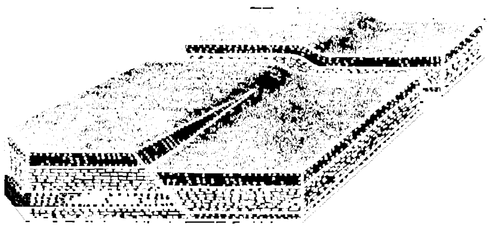


FIG. 249. A hinge fault passing into a monoclinal fold. Front of block about 1 mile long.

A sunken tract due to downfaulting, or to uplift of adjacent areas, is called a *graben* (German for trough or ditch). Illustrations are the Jordan Valley and the Dead Sea basin in Palestine, and the great rift valleys of Africa. An upstanding mass between two faults is a ***horst*** (Fig. 291).

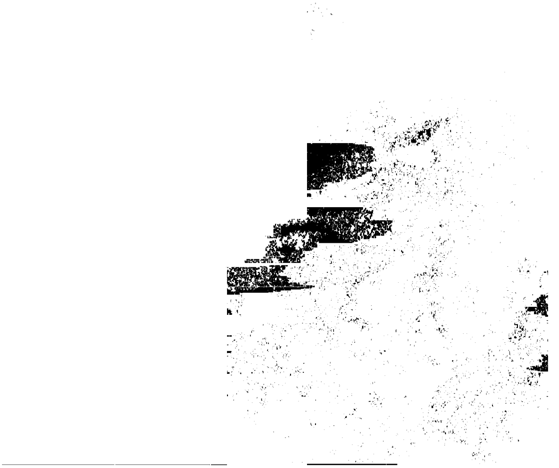
Many of the great fractures mentioned above are normal faults. Detailed study of eroded mountain areas has disclosed reverse faults of enormous displacement, some of them dipping less than 20 degrees, or even horizontal. Such faults, which occur only in regions where folding and crushing of the rocks have been exceptionally severe, are commonly known as *thrust* faults, or simply *thrusts.* Many of these are of such magnitude and importance that they have come to be considered as a special class of faults. The surface on which movement has occurred is spoken of as the thrust surface, or less accurately as the thrust plane.

Such thrusts have been discovered in the Alps, in northwestern Scotland, in Scandinavia, in the southern Appalachians, in the Rocky Mountains, and in many other mountain regions. The horizontal displacement of lower, older formations over younger rocks ranges from a few miles up to more than 25 miles for individual thrusts (Figs. 301–306, p. 466). In many places erosion has partly destroyed the comparatively thin plate of the older rocks above the thrust, leaving isolated remnants resting on much younger foundations. An excellent example is Chief Mountain in Glacier National Park, Montana (Fig. 307, p. 467). Such isolated masses that obviously are far removed from their original positions are called "mountains without roots."

The deciphering of these great thrusts is one of the triumphs of modern geologic research. It has led to rational interpretation of mountain structure that before seemed illogical and chaotic. An illustration of the practical value of such research was the discovery in Belgium of important deposits of anthracite coal after it was recognized that the coal-bearing formations lie concealed beneath older rocks that were pushed over them in the development of a great flat thrust.

Surface Expression of Faults. Displacement of the Earth's surface by a normal or a reverse fault gives rise to a cliff or scarp (Fig. 250). Within historic time, movements that caused severe earthquakes have resulted in new scarps from a few feet to nearly 50 feet high (Fig 254) Many other cliffs and mountain fronts, some of them hundreds or thousands of feet in height, are recognized as fault scarps, but it is extremely improbable that any one of these was formed by a single displacement. There is strong evidence that the stress responsible for slipping on a fault is relieved temporarily by abrupt displacement of a few feet or tens of feet, and then accumulates for years or even hundreds of years before the movement is repeated. Thus every high fault scarp is a very old feature judged by human standards, although it may be youthful from a geologic viewpoint.

Such a scarp made directly by faulting is called simply a fault scarp (Fig. 243). Erosion starts to modify it as soon as it begins to be



ROYAL CANADIAN AIR FORCE.

Fig. 250. Looking southwest along great fault at McDonald Lake, near Great Slave Lake, Canada. Maximum height of scarp is about 900 feet. This fault can be traced for more than 300 miles.

formed, and the upper portion of every high scarp is considerably dissected, both because that part has been subject to erosion since the scarp was first initiated, and because erosion is most effective at high altitudes (Fig. 292, p. 456). After the faulting has ceased the entire scarp is continuously lowered, and it slowly retreats from the position of the fault. As erosion progresses, the scarp passes through the stages of youth, maturity, and old age, and finally, on peneplanation of the region, differences in elevation on opposite sides of the fault almost or entirely disappear.

A large fault displacement as a rule brings together on opposite sides of the fault rocks that differ materially in composition and resistance. As erosion proceeds the original scarp (Fig. 251) is eventually destroyed, but as long as the entire region has considerable altitude the position of the fault is marked by a cliff or steep slope, because weak rock on one side is removed faster than resistant rock on the other. A scarp that is due entirely to differential erosion along a fault is called a *fault-line scarp*, to distinguish it from the earlier scarp that resulted directly from the faulting movement (Fig. 252). Since resistant rocks at any particular level are as likely to occur on the downthrow as on the upthrow side of the fault, not uncommonly the block that originally stood the higher is eroded far below the other, and the resulting fault-line scarp faces in a direction opposite that of the earlier scarp (Fig. 253).

Fig. 251

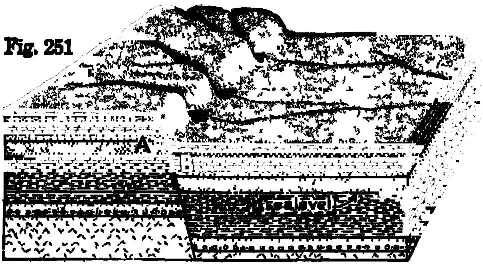


Fig. 252

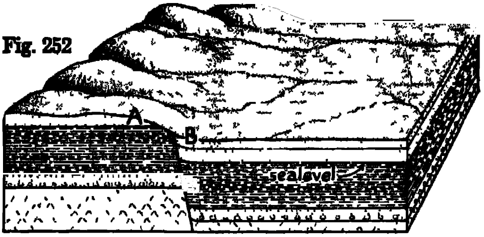
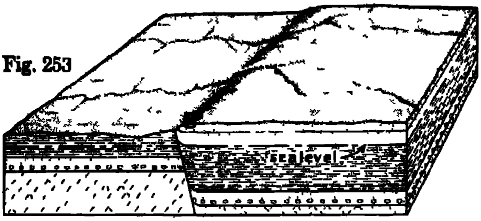


Fig. 253



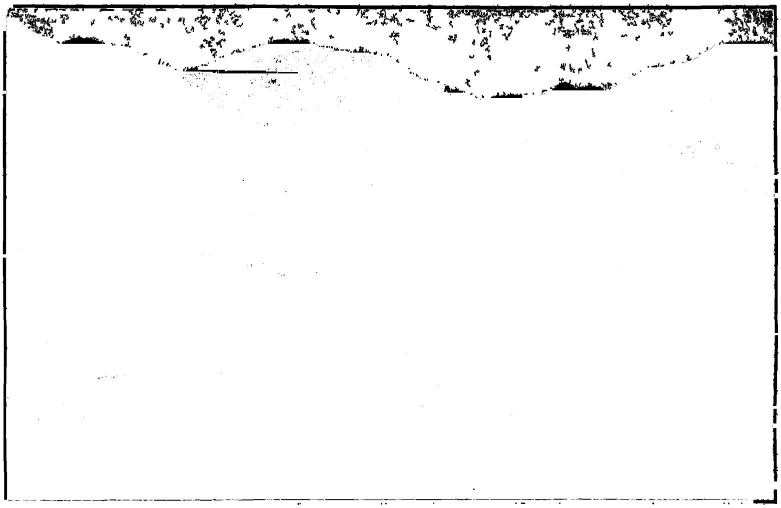
Figs. 251–253. Ideal development of a fault scarp through progressive erosion

Fig. 251. Scarp made directly by faulting. Only a little of the topmost stratum has been eroded from the uplifted block, and streams are cutting deep valleys into the high mass, spreading the debris on the downthrown block.

Fig. 252. The same fault shown in Fig. 251, at a later stage in the cycle. *A* and *B*, points at the top and base of the scarp at this stage, were beneath the surface when the initial scarp existed (Fig. 251). Therefore the original scarp has been entirely destroyed, and the scarp at this later stage is wholly the work of erosion; it is a *fault-line scarp*

Fig. 253. A possible development in a late stage of the cycle. After the sandstone layer at *A* (Fig. 252) is destroyed the weak shales beneath are eroded rapidly. The sandstones beneath *B*, on the downthrown block, make erosion on that side more difficult Aided by favorable changes in drainage, these differences in the bedrock may result in a fault-line scarp facing opposite the original fault scarp.

Some fault-line scarps are developed by regional uplift after peneplanation has removed the differences in elevation on opposite sides of a fault (Figs. 255–257). There are excellent illustrations in central Connecticut and Massachusetts, where scarps destroyed by the erosion that formed the New England peneplane have been renewed since the region was warped up during the Cenozoic era. These examples show the necessity of careful geological study in determining the exact history of any fault.

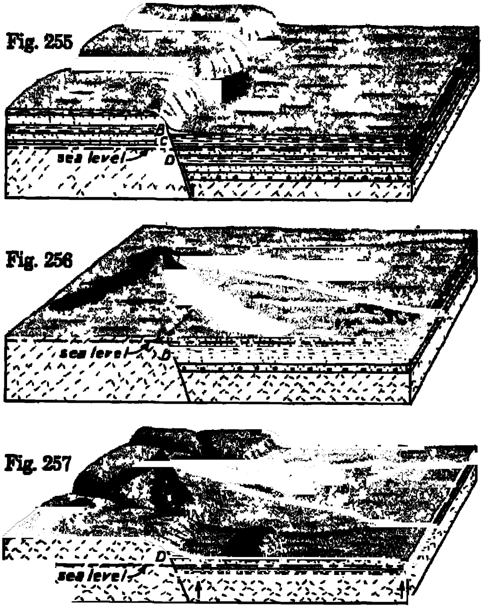


Eliot Blackwelder.

FIG. 254 Result of recent faulting at the base of the Sonoma Range, Nevada The new scarp, appearing as an irregular band extending across the view at the top of the alluvial slope, is about 20 feet high. The photograph was taken from a point 3 miles away

In southwestern United States, where large crustal movements have occurred in late geologic time, and where climatic conditions favor the development of excellent outcrops, there are many fine examples of fault scarps as well as fault-line scarps The east slope of the Sierra Nevada and the west slope of the Wasatch Mountains are fault scarps that are youthful in a geologic sense At the base of each of these ranges some of the movement has occurred so recently that scarps exist almost uneroded in the fans of weak alluvial material brought down by the streams (Fig. 254). The last recorded movement along the east base of the Sierra Nevada took place in 1872 These faults have [illegible] by many successive displacements, and the upper part of each [illegible] been greatly eroded. Some of the great faults in the Plateau region [illegible]

Arizona and Utah, through which the Colorado River cuts its way, are marked by prominent cliffs. These cliffs have been described as actual fault scarps; but most of them are fault-line scarps developed after advanced erosion and renewed regional uplift.



Figs. 255–257. Development of a fault-line scarp during two cycles of erosion. Length of block about 5 miles.

Fig. 255. A fault scarp (*A–B*), modified by erosion but still growing by downward movement of the right-hand block.

Fig. 256. The surface is reduced to a peneplane, through the point *C*, and lava flows spread across the position of the fault.

Fig. 257. The entire region is uplifted uniformly, as indicated by arrows, and erosion forms a new scarp (*C–D*) by more rapid removal of the weak sediments on the downthrown block. Presence of the lava flow, crossing the fault unbroken, proves that this is a fault-line scarp, developed wholly by erosion and not by renewed movement on the fault.

Many old faults with vertical displacements amounting to thousands of feet are practically unrecognizable in the present topography. Obviously this relation in each case indicates erosion of great magnitude. Either there was a high fault scarp that slowly wasted away, or the growth of the displacement was so slow that erosion almost kept up with it. This last suggestion is not unreasonable, for we have evidence of repeated movements on some faults after intervals of long geologic periods. If individual movements are small and widely separated in time, or if the rocks on the upthrow side of a fault are exceptionally weak, a prominent scarp has no chance to develop.

Economic Importance of Faults. Faults and other fractures have commonly been the channels along which ore solutions have found their way and deposited valuable metallic minerals (p. 531). The famous copper-bearing veins at Butte, Montana, were formed in successive stages along sets of great faults that intersect and offset each other to form a baffling complex of blocks. Efficient mining of these broken veins requires intensive study of the faults, to determine the order of their formation, directions and amounts of displacement, and other facts necessary for guiding the miners in their expensive tunneling operations. The "losing" of a vein where it has been offset and thus hidden by faulting is one of the commonest difficulties encountered in mining.

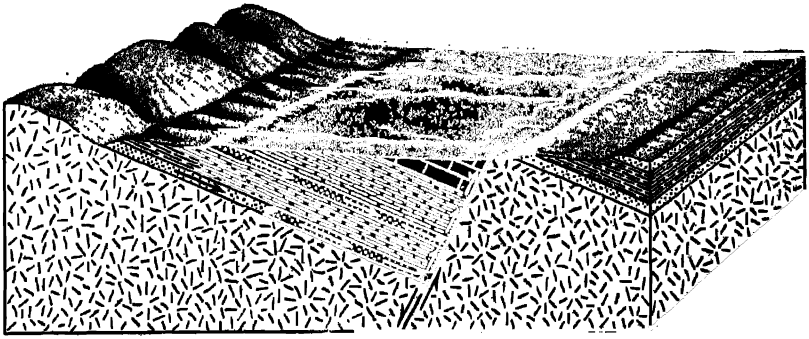


FIG. 258. A line of springs along a fault. The diagram represents an area about 20 miles long. After the strata were faulted and tilted, erosion reduced the upthrown block (at right of fault), and the present drainage became established. Artesian conditions (p. 123) exist in two sandstone formations; water rises from these aquifers along the fault (see full arrows), issuing as springs at the most favorable points. Since the granite at the right of the fault is almost impermeable, nearly all the water in the two aquifers is diverted to the Earth's surface. Springs of this type are called fault-dam springs. (Direction of fault movement indicated by half-arrows.)

The occurrence and recovery of subsurface water, one of our greatest and most vital natural resources, are controlled to an important degree by geologic structure. Faults cutting across some artesian systems have brought impermeable rocks across the faulted edges of important aquifers, thus diverting the flow of water to form artesian springs (Fig. 258).

Engineers responsible for the construction of dams, aqueducts, and other large structures seek to avoid locations on or near faults, particularly those that are known or suspected to be still active. Shattered bedrock adjacent to "dead" faults may be too weak and permeable for satisfactory damsites. Movement along an active fault beneath or near a dam or aqueduct might cause serious damage.

CAUSE OF WARPING, FOLDING, AND FAULTING

The immediate cause of rock deformation is comparatively simple and generally agreed upon: it results from stresses set up in the outer shell of the Earth. Some rocks yield by warping or folding but others, too massive or too brittle to permit yielding of this kind, become fractured. Compressive stresses give rise to folds and also to reverse faults and thrusts. In regions of broad warping, perhaps in connection with isostatic adjustment (p. 17), differential vertical stresses may produce normal faults. Thus over wide regions where the strata are not otherwise disturbed, as in parts of the Colorado Plateau, they are penetrated by steep fractures on which there have been great displacements.

The *ultimate cause* of faulting and folding evidently depends on those processes within the Earth which give rise to compressional and tensional forces and so set up stresses in the lithosphere. The forces themselves are hidden and can be inferred only from their effects. As the subject is obscure at best, and speculation must be guided by consideration of all available facts, it is best to postpone inquiry into the ultimate cause of crustal deformation until the structure and history of mountains have been discussed (Chap. 19).

UNCONFORMITY, A RECORD OF ANCIENT DIASTROPHISM

Sedimentary deposits, especially those laid down on sea floors and later elevated above sealevel, give a faithful record of the history of their times. In general the persistence of nearly uniform conditions, without any disturbance by crustal movements, results in continuous deposition of parallel strata with similar composition or with gradual change from one type of material to another, as from sandstone to shale. The strata in such a series are said to be *conformable* with each other. But if crustal movement causes uplift and erosion of sedimentary beds, any later series of strata deposited above them is *unconformable* with them (Fig. 259). The *unconformity* of one group of rocks with another is evidence of a definite succession of events and therefore is a significant part of the geologic record.

The most conspicuous unconformities result when eroded mountain regions, with their truncated folds and steeply tilted sections of sedimentary strata, are buried by later sediments. At practically every point the edges of the older beds form a considerable angle with layers in the later series, and therefore the term *angular unconformity* is appropriate (Fig. 259). Two or more angular unconformities in the same section indicate two or more crustal disturbances, each followed by long-continued erosion and a later period of quiet when sediments accumulated.

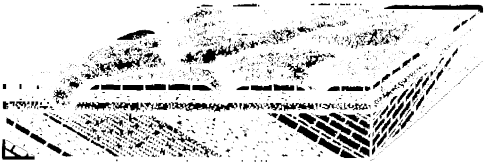


FIG. 259. Angular unconformity between two series of marine strata. The lower series was deposited as a conformable sequence, then tilted and eroded; later submergence allowed deposition of the upper series, after which the area was lifted above sealevel and the present drainage system was developed. The area represented covers several square miles.

On the other hand, suppose a wide shallow-sea basin is warped up almost uniformly to make a land surface. Erosion sets in, and a considerable thickness of the marine strata is stripped away over the entire area. If the sea again invades the district, new strata are deposited above the old erosion surface, and a later upwarping, with consequent cutting of stream valleys, brings to view the older as well as the younger rocks (Fig. 260). Because the older strata experienced no appreciable tilting they are essentially parallel to the beds above them; but the two series are separated by the old buried erosion surface. The unconformity between the two groups of strata is important, because it represents crustal movement of wide extent and a long time interval during which no sedimentary record was being formed in the region, while part of the earlier record was being destroyed. However, this

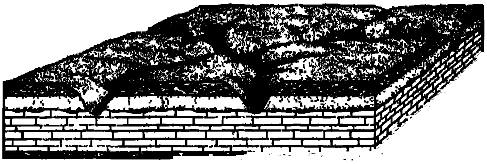


FIG. 260. Disconformity between two series of formations. No tilting has occurred; but the lower series was uplifted uniformly and eroded before deposition of the upper strata. Compare Fig. 259.

type of unconformity is not so conspicuous as an angular unconformity, which attracts attention at once by the divergence of beds above and below. In view of this striking difference, the special name *disconformity* is applied to an unconformity that is not characterized by any angular divergence between the two groups of strata.

It is difficult to recognize some disconformities, particularly if the same type of rock lies above and below. This is true in parts of the Mississippi Valley, where limestones deposited during several geologic periods are almost as flat as the old sea floor on which they were formed. Geologists have learned to recognize and date the different formations by careful study of the fossils contained in them. For some sections exposed in quarries or in natural cliffs it can be stated confidently that a particular layer is many millions of years younger than the layer directly beneath it, although the two are parallel and superficially much alike. Conditions were exceptional during much of geologic time in what is now the Mississippi Valley. During long geologic periods shallow seas covered much of the region. Since land was far distant, only limy deposits accumulated on much of the sea floor. Slow warping movements made the sea now shallower, now deeper, and large tracts lay for a time above sealevel, but so low that very little was eroded from the surface and no hill-and-valley topography could be formed. Other tracts were covered with water so shallow that the breaking of the waves on the bottom prevented any permanent deposition for long intervals of time. When enough deepening finally occurred to permit renewed sedimentation, the new strata were limestones, much like those directly beneath. In this way several disconformities were developed which, although they are inconspicuous, represent long intervals of time.

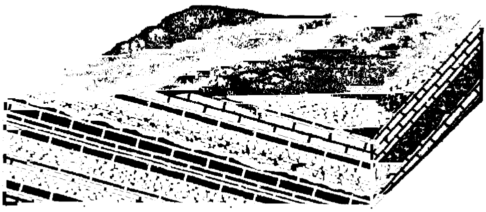


FIG. 261. Disconformity between two series of strata, both series tilted together. A large part of the upper series was destroyed by stream erosion in developing the present surface after the tilting occurred. Compare Fig. 260.

A distinguishing characteristic of disconformities is the parallelism of the younger and older strata; it is not essential that the beds shall be horizontal or undeformed. Commonly there has been disturbance, either slight or severe, at some time after the younger formation was deposited. Since both groups of beds are deformed together they remain parallel (Fig. 261).

Unconformities Involving Nonsedimentary Rocks. Although the conception of unconformity is inseparable from sedimentary processes, the old surface on which strata are unconformably laid may be developed on rocks of any kind. Granite and other massive igneous rocks, as well as metamorphic rocks of all kinds, have been bared by erosion in wide areas and later covered with sediments (Fig. 262). The contact is a surface of unconformity if the sediments were deposited on it. If the contact between rock masses of unlike character results from faulting or from igneous intrusion, however, the term unconformity does not apply.

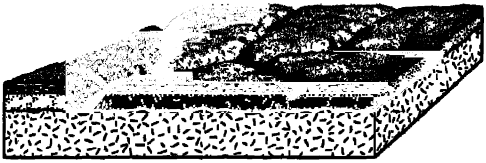


FIG. 262. Unconformity made by erosion of granite bedrock and later deposition of sedimentary strata on the irregular surface.

Common Features of Unconformities. Since every unconformity involves some degree of erosion before deposition of the younger sedimentary series, the surface at the base of this series is more or less irregular. If the old land surface on which deposition began was hilly, the buried hills and valleys can be recognized provided the unconformity is well exposed. Generally the old lands were worn down to low relief before any permanent deposits were formed. If the unconformity represents invasion of the land by the sea, it is probable that submergence occurred very slowly and wave action near the advancing shoreline had opportunity to plane down the most prominent irregularities. If the sediments on the surface of unconformity are ordinary stream deposits, probably the old land had been reduced by streams to a peneplane, because most sediments laid down on a high, rugged land mass are swept away by continued erosion. Exceptions are the deposits in basins of interior drainage, which accumulate until they bury hills and mountain ranges (p. 115). Through changes in climate or important shifts in drainage some ancient basin deposits of this kind have become deeply dissected, and the extremely irregular contact between the sediments and the rugged old surface is exposed. Tributaries of the

Colorado River have exhumed old buried mountain topography of this sort in southern Nevada.

Usually the sediments deposited on a surface of unconformity are distinctly different in character from any sedimentary formations beneath. Whenever the sea invades the land every part of the submerged area is passed over by the advancing shore zone, with its characteristic deposits of pebbles and coarse sand (p. 279; Figs. 190, 191). As a common result a *basal conglomerate,* composed in part of the beach materials, lies directly above the old land surface.

Whether the first sediments above an unconformity were deposited by the sea or by streams, some pebbles in the lowest layer had their source in the rocks on which the deposits rest. The presence in a conglomerate of pebbles obviously derived from granite beneath proves clearly that the granite is older than the overlying strata and therefore is not intrusive into them.

Unconformities and Earth History. A study of unconformities emphasizes the close relationship between crustal movements, erosion, and sedimentation. The lands are being worn down and sediments are accumulating continuously. But slowly the scenes shift. Folding or faulting in a geosynclinal basin, or upwarping at the edge of a continent, brings to light the strata built up during former ages. At once the forces of erosion attack the strata of shale, sandstone, and limestone, and in the act of destroying them reveal the record they contain. In the deeply incised stream valleys we see angular unconformities that testify to severe folding in remote geologic periods, or widespread disconformities that indicate ancient upwarping of large areas. As erosion bares these secrets, the streams carry away detritus and deposit it, perhaps on a land surface recently submerged. Thus construction in one place supplements destruction in another; the whole complex record shows the constant struggle between deep-seated forces that produce the major relief features and processes at the surface that strive to keep the lands featureless and low.

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Chapter 16

EARTHQUAKES

Nature and Origin

Sensitive modern instruments distributed widely on all the continents show that the outer part of the Earth trembles almost constantly. Most of the tremors are not perceptible to our senses, although they are recorded instrumentally; many of them are caused by various activities at the Earth's surface, such as the ebb and flow of the tides, changes in atmospheric pressure, the rush of traffic on city streets, and the tumbling of streams over high falls. But numerous tremors, both powerful and weak, are the result of disturbances within the body of the Earth itself and logically are called *earthquakes.*

Human Interest of Earthquakes. The scientific value of earthquake study lies in the large amount of information it yields concerning the interior of the Earth. There is a more immediate interest, however, in the relation of earthquakes to human affairs. From the earliest recorded times the recurrent shaking of the ground, with consequent destruction on the surface, has been a cause of terror to mankind. Repeatedly, in numerous localities, populous communities have suffered great loss of life and property. Destructive earthquakes recorded during the brief span of written human history are numbered in thousands. Geologic evidence indicates that violent shocks have been recurrent throughout the history of the Earth; and there is every reason to expect their frequent occurrence in the future.

The serious aspect of earthquakes from the human viewpoint is realized on review of some major catastrophes. On September 1, 1923, more than 140,000 lives were lost as a result of the Tokyo earthquake, and the estimated property loss was nearly $3,000,000,000. The shocks at Messina in 1908 and at Kansu, China, in 1920 were equally disastrous to life. According to written records, more than a million and a half persons were killed in China by ten shocks between the eleventh and twentieth centuries; and Mallet, a profound student of earthquakes, estimated that for the whole Earth at least 13 million lives were lost through destructive shocks in the course of 4000 years. Although activities of man himself, such as the waging of war and the operation of automobiles, result in a much higher death rate, nevertheless earthquakes are especially productive of fear, probably in part because they come without warning, and in part because to most people their origin seems mysterious. Study of earthquakes from the viewpoint of geology and physics has dispelled a part of the mystery. The study is called *seismology* (sīs-mŏl′o-jĭ) from the Greek *seismos,* earthquake.

Causes of Earthquakes. An earthquake is a trembling or undulatory motion in the elastic rocky shell of the Earth, communicated to it by sudden jarring of some kind, just as a bell is set in vibration by a smart tap on its side. The jarring impulse evidently is the immediate cause of the earthquake; but what is the origin of such disturbances? Ancient philosophers tried to answer this question, but their speculations were without scientific basis. Aristotle taught that subterranean cavities are filled with air which in its struggles to escape causes the ground to shake. Lucretius, in his *De Rerum Natura* (first century B.C.), made a shrewder guess in assuming roof collapse in vast caverns as one of the primary causes. Probably some local shocks have originated in this way in karst regions such as those of Yugoslavia and Kentucky (p. 137).

Volcanic Earthquakes. It is well established that some earthquakes are associated with volcanic activity. The violent outbursts of Krakatoa in 1883 (p. 327), of Bandaisan (Japan) in 1888, and of Paricutin (Mexico) beginning in 1943, were accompanied by shocks that were severe locally; but most earthquakes of this class are of low intensity and affect limited areas. Moreover many outbursts are not attended by any shocks, or at best are accompanied by only feeble tremblings, such as occurred during the eruption of Mont Pelée in 1902. For a long time it was thought that volcanic action was an important source of earthquakes, and this idea still appeals to popular fancy; but careful comparison of the two phenomena, especially in Japan, has shown that there is no necessary connection in occurrence between heavy earthquakes and volcanic eruptions. Instruments near Kilauea, in Hawaii, record frequent minor tremors—sometimes hundreds of them in a single month. Very few of these are accompanied by visible volcanic activity, although it is probable that shifting of magma at some depth is the principal cause of the local shocks.

Crustal Movements the Chief Cause. Most of the major earthquakes result from sudden yielding to stress in the Earth's crust, either by formation of new fractures or by abrupt displacement along the walls of existing faults (p. 377). In many areas visited by disastrous shocks the surface of the ground has been broken along faults, and the amount of displacement is clearly indicated. Commonly these movements take place along old fault zones which bear the marks of repeated displacement. In California a great fracture zone can be followed almost continuously, by means of its peculiar surface expression, from the southern part of the state northwestward for 600 miles. This feature, the San Andreas fault, passes near the city of San Francisco (Fig. 263). On April 18, 1906, abrupt movement along at least 270 miles of this fracture caused a destructive earthquake. The length of this break is exceptional among recorded fault movements; but similar breaks 25 to 50 miles long are not uncommon.

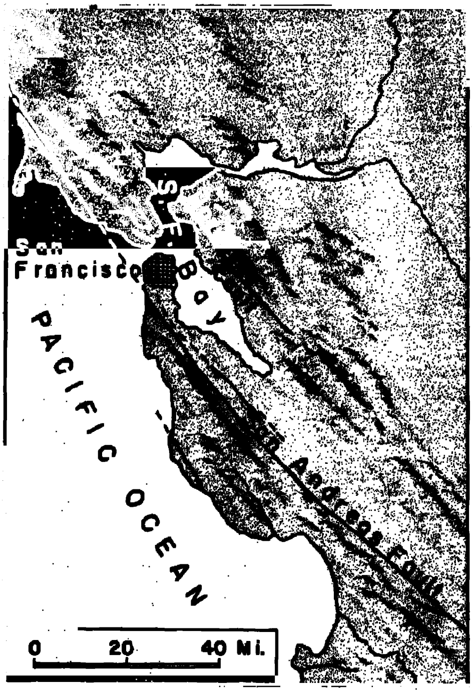


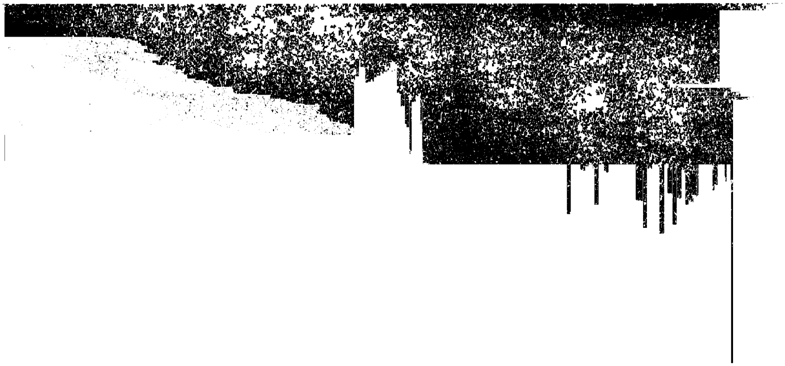
FIG. 263. The San Andreas fault and other faults in the vicinity of San Francisco, California. The mountain ridges are a part of the California Coast Ranges.

A detailed study along the San Andreas fault, after the rupture in 1906, yielded valuable information on the nature and amount of the displacement. In most places no scarp was made by the faulting, because the motion was almost entirely horizontal, parallel with the fault (Fig. 264). This fact was established beyond question by the offsetting of roads, fences, and other features that extended across the break. The largest measured displacement was 21 feet. On May 18, 1940, comparable movement occurred in extreme southern California, along a fault that appears to be a branch of the great San Andreas fault, causing a severe earthquake in Imperial Valley (Fig. 265). More commonly a part of the motion on a break of this kind is vertical and forms a steep scarp (Fig. 254, p. 383). The great Yakutat Bay (Alaska) earthquake in 1899 resulted from an abrupt vertical displacement that lifted part of the coastal belt nearly 50 feet.

When we consider that the walls of a fault are pressed closely together, that movement is possible only by overcoming great frictional resistance, and that the displacement, once it occurs, takes place almost instantaneously, it is not surprising that powerful vibration is set up in the adjacent rock. The exact nature of movement along the San Andreas fault in 1906 was made the subject of special study, and it was

M. S. KENNEDY.

Fig 264. Air view along the San Andreas fault zone, at the west base of the Temblor Range, California. The position of the fault is marked by the nearly straight furrow between the hills and the low ground. View taken from altitude of 6000 feet.

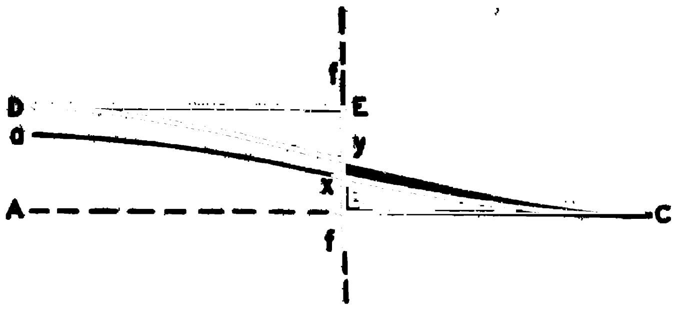


concluded that slow movement of great crustal blocks on opposite sides of the fault had been in progress for many years before the earthquake occurred. Deformation in the rock was by slow bending, until the strain could be borne no longer and relief occurred by *elastic rebound* along the old fracture (Fig. 266). According to this view the sudden movement of 1906 was confined to a comparatively narrow belt closely adjacent to the fault and did not involve the immediate shifting of great segments of the crust. The distorted rocks along the fault straightened with the abruptness of the springing of a steel trap; energy that had been accumulating for decades was released in an instant, striking the Earth a sharp blow that made it tremble. Vertical displacement on a fault is supposed to occur by similar elastic rebound after slow bending (Fig. 267).

A visible fault does not appear in every area visited by an earthquake. Commonly direct evidence of crustal movement is wanting, especially in connection with mild or moderate shocks. It is believed that actual displacement occurs frequently at considerable depth and does not reach the surface. This is a logical inference, since every break must be limited in extent, vertically as well as horizontally. Earthquakes of the first rank, however, are generally restricted to regions of active faulting, for which there is evidence at the surface. Many major shocks originate under the sea, and in some places soundings made after such an earthquake have demonstrated that the sea floor was displaced

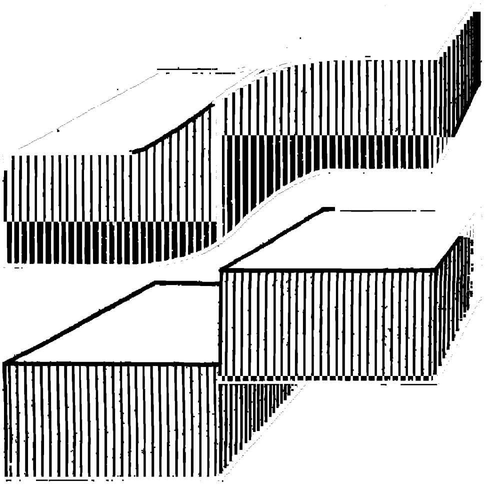
HETZEL, EL CENTRO, CALIFORNIA.

Fig. 265. Horizontal displacement at the time of the Imperial Valley (California) earthquake of May 18, 1940. Before the movement the bushes in the left foreground were directly in line with the row of bushes marked by the man. Position of fault is indicated by the disturbed ground.



After A. C. Law

FIG. 266. Illustration of the elastic rebound theory of earthquakes. Sketch map of an area in the vicinity of the San Andreas fault (*f–f*). *ABC*, position of a line (say a fence) directly before strain began to build up in the fault zone; *axC*, position of the line several years later, when the rocks had been considerably deformed; *DyC*, position of the line immediately before the movement of 1906; *DE*, *BC*, the two offset portions of the line after slipping occurred on the fault, allowing *Dy* and *yC* to straighten. *C* is a stationary point outside the zone of movement. (Scale of *BE* greatly exaggerated; this distance is only about 20 feet, whereas *AC* is 200 miles or more.)

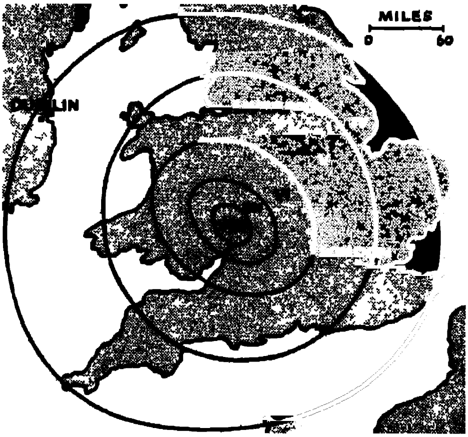


After A. Sieberg.

FIG. 267. A segment of the crust showing elastic rebound from vertical strain, with the formation of a fault scarp. The upper block shows bending (greatly exaggerated) by accumulating strain before movement; the lower block represents the displacement after the elastic rocks have sprung into a position of no strain.

at the time of the shock. Transoceanic cables have been broken by abrupt displacement of the floor athwart their courses.

Intensity of Earthquakes. Earthquakes vary in strength from minute tremors to the irresistible shocks that cause major human disasters. In order to compare the effects of different shocks, and the results of the same shock in different places, an arbitrary scale of intensity has been adopted; on this scale earthquakes are classified from I to X in increasing order of strength. A shock of intensity I is hardly noticed even by an experienced observer, but is recorded by delicate instruments; one of intensity V is felt by nearly everyone awake, disturbs furniture in houses, and rings church bells; intensity as high as VIII is required to throw down chimneys and crack the walls of brick or stone buildings; intensity X causes widespread destruction of buildings and disturbance of the ground.

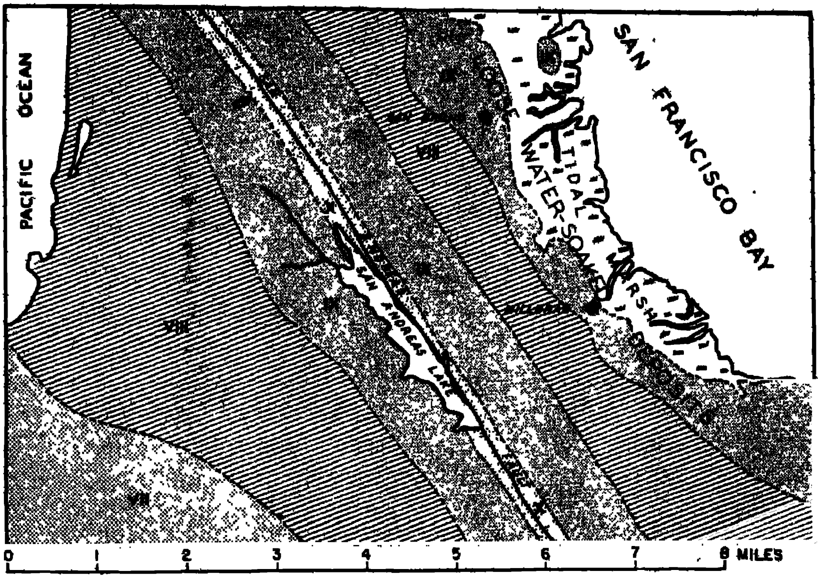


Modified from Charles Davison.

FIG. 268. Isoseismal map of an earthquake with epicenter near Hereford, England, in 1896. Roman numerals indicate intensity in the several zones. Note that the isoseismal lines are not circular, but outline oval-shaped areas elongate northwest-southeast.

By the use of such a scale a region affected by an earthquake is divided into intensity zones, and the lines separating adjacent zones are called *isoseismals*. Ideally they inclose nearly circular or elliptical belts around the area of highest intensity (Fig. 268); but for an earthquake caused by slipping on a long fault, like the California earthquake of 1906, the isoseismals define narrow belts that are nearly straight for long distances (Fig. 269). Differences in the nature of the bedrock and also the irregular distribution of unconsolidated sediments cause pronounced local irregularities in isoseismal maps (Fig. 269).

The scale of intensity based on observed destructive effects is useful for comparative purposes; but at best it gives only a crude and qualitative conception of the energy displayed by an earthquake at each locality affected. A more scientific scale, in use by some special students of earthquakes, is based on comparison of graphic records made by an instrument of standard type at a distance of 100 kilometers from an earthquake center. Engineers interested in building earthquake-resisting structures have devised another instrument that gives precise records of intensity, expressed in terms of the acceleration of a mass set in motion by an earthquake shock. As the practical study of earthquakes continues, no doubt these instruments will be further perfected and will be adopted for general use in seismic regions.



California Earthquake Commission.

FIG. 269. Zones of intensity in the lower part of the San Francisco peninsula, for the earthquake of 1906. Roman numerals indicate degrees of intensity in the belts concerned. The highest intensity (X) was restricted to a narrow belt on each side of the fault, and in general there was a gradual decrease on both sides; but the belt of loose deposits near the Bay caused abnormal intensity. Since there were no dwellings or other works of man on the tidal marsh, the intensity in that area is not known.

DISTRIBUTION

Seismic Belts. Although earthquakes occur in all parts of the world, those that occur on land are concentrated in certain well-defined tracts, most of which lie in two great *seismic belts*. One of these follows the western coast of North and South America, the Aleutian Islands, and the island groups along the eastern coast of Asia such as Japan and the Philippines, and thus borders the Pacific Ocean on the east, north, and west. The other includes the Mediterranean, the Alps, the Caucasus, and the Himalayas and continues into the East Indies, where it intersects the first belt (Fig. 217, p. 332). In a general way these zones coincide with the great volcanic belts (p. 331); and this fact might appear to support the idea that volcanoes are an important cause of earthquakes. However, since the belts correspond closely to young mountain systems and other marks of recent crustal movement, it is probable that earthquakes and volcanoes have a common relation to this disturbance of the crust. It is a notable fact that where the seismic belts lie directly along the continental borders, as on the coast of Chile and the eastern coast of Japan, the edge of the continental mass descends rather abruptly, without any broad intervening shelf, to great depths. These deeps (Fig. 2, p. 4) are great troughs that appear to be sinking while the bordering lands are rising. We conclude that these are zones of weakness in the Earth's crust where stresses are being relieved by frequent movements, and in which therefore earthquakes recur at short intervals.

It is commonly thought that certain regions are practically exempt from danger of earthquakes because no real disaster has happened in them since they have been settled and cities have sprung up within them. It is true that most of the Atlantic coasts, and large areas in continental interiors, are relatively free from earthquakes. The comparative stability of the Atlantic seaboard as compared with that around the Pacific is emphasized not only by the historical record, but by the existence of a wide continental shelf, which is in strong contrast with the deep water near the Pacific coasts (p. 4). However, the experiences of the central Mississippi Valley in 1811 and of Charleston (South Carolina) in 1886 warn us that no locality may be entirely exempt. Even in New England, which is not recognized as a seismic tract, there has been an average of one perceptible tremor a year since the settlement of the country. Probably the only one of these that approached maximum intensity occurred in 1755, with its center near Cambridge, Massachusetts.

Submarine Shocks. The location of seismic belts near the margins of continents suggests that many earthquakes are of submarine origin. Their occurrence beneath the sea is shown directly by shocks communicated to vessels on the surface above, and by rupturing of submarine cables. Since the invention of sensitive instruments by which it is now possible to record distant earthquakes and determine their location, it has been learned that a large majority of earthquakes occur on the floor of the Pacific. A prominent seismic belt on the floor of the Atlantic coincides rather accurately with the Mid-Atlantic ridge (Fig. 217, p. 332). The most conspicuous mark of a submarine earthquake is the huge wave that commonly is generated in the sea by disturbance of the floor. Such waves have long been known as "tidal waves," a misleading name since they have no connection with the tides. They are now generally known to seismologists by their Japanese name, *tsunamis;* they are also called seismic sea waves. Some are of immense size, measuring 100 or even 200 miles from crest to crest, and as much as 40 feet in height. They are so broad that in the open sea they are not ordinarily perceived; but on approaching a coast those of large size pile up in huge breakers and, sweeping far inland, cause enormous damage and loss of life.

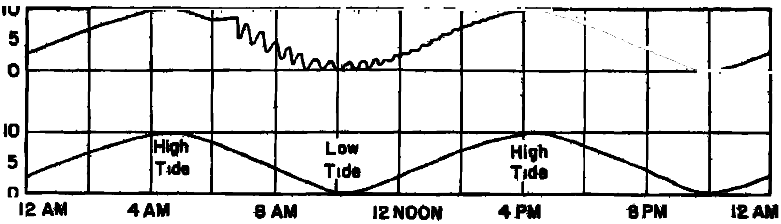
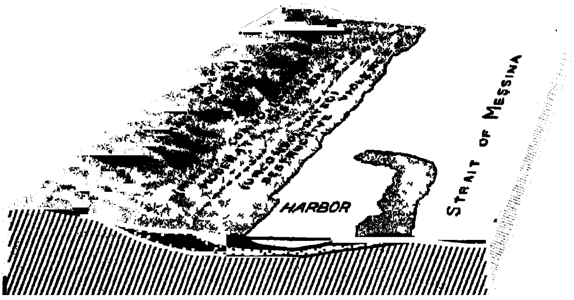


FIG. 270. Records by a tide gage. The lower curve shows the normal succession of high and low tide, the upper curve shows rapid oscillations, with a period of about 25 minutes, caused by a seismic sea wave superposed on the tidal record. Horizontal spaces record equal time intervals as the paper is moved uniformly by clockwork. Vertical spaces show heights in feet as recorded by the rising and falling pencil of the gage.

Lisbon in 1755, Japan in 1703 and 1896, and Peru in 1868, suffered from great and disastrous tsunamis. In April, 1946, a tsunami that originated in the Aleutian Deep caused widespread destruction in the low coastal areas of the Hawaiian Islands 2000 miles away. The number of victims of a single inundation of this kind has been as great as 100,000. These vast waves are felt over whole oceans and move with tremendous speeds, ranging from 300 to 500 miles per hour. Some that originated near Japan have crossed the Pacific in about 12 hours. At such distances their height may be only a few inches; but they ebb and flow like small tides in periods of 15 to 30 minutes, and these variations are registered as undulatory lines on the record of a tide gage (Fig. 270).

SEISMOLOGY

Transmission of Earthquake Vibrations. It is important that we distinguish clearly between cause and effect in earthquake phenomena.



Modified from A. Sieberg.

FIG. 271. Correspondence between belts of intensity and geologic formations at Messina, Italy, in 1908. The water-soaked loose sediments appear in black in the vertical section on the front of the block.

The displacements shown in Figs. 265 and 267 are not, as is commonly supposed, the *results* of earthquakes; they represent the *causes*. The effect of the sudden movement along a fault is to set up vibrations that move outward from that place, as circular waves spread out from the position at which a stone strikes the surface of a pond. Thus the earthquake is propagated as a series of waves in the highly elastic body of the Earth. In the passage of these waves the particles of bedrock usually are displaced only a fraction of an inch, even near the origin of the shock; but loose objects on the surface may be thrown several feet by the impulse. This effect is illustrated in miniature by placing a pebble or similar object on a board floor and striking the floor a sharp blow with a hammer; the pebble can be made to leap several inches although the floor in transmitting the impulse does not move perceptibly.

Oscillation of the ground during an earthquake reaches a maximum in thick deposits of recent sediments that are saturated with water. At San Francisco the devastation was most acute on the low flat near the bay, which is underlain by loose, water-soaked silt and sand (Fig. 269); well-constructed buildings on bedrock, even much nearer the fault, suffered less damage. At Messina, Sicily, in 1908, there was an even clearer correlation between maximum destruction and recent sedimentary deposits (Fig. 271). The contrast in behavior of unconsolidated sediments and of bedrock may be illustrated by striking sharply the side of a bowl containing jelly. The bowl is set vibrating with resulting sound, but actual motion in the walls of the vessel is imperceptible. However, the same impulse transmitted to the jelly sets up longer waves that are visible. Excessive destruction on saturated loose sediments is caused by this larger wave motion as compared with the behavior of solid rock. Observers at Charleston, South Carolina, during the destructive earthquake of 1886, reported that they saw the ground rise and sink with the swift passage of large swells.

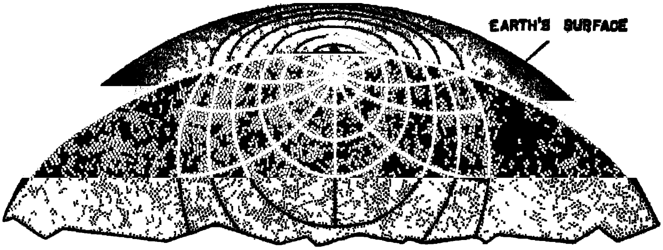


FIG. 272. Ideal section through part of the Earth, to illustrate the spreading of elastic waves in nearly spherical form away from an earthquake focus (the inner small circle). These successive wave fronts intersect the surface of the Earth, as indicated by the lines at the top of the diagram. Vertical path emerges at the epicenter, *e*. The impulse travels more rapidly with depth, because of the increasing pressure; hence the wave fronts are not truly spherical. Any line (except the vertical) along which the impulse may be traced from the focus to a point on the Earth's surface is curved downward, as shown; such a line is everywhere at right angles to the advancing wave front.

If we think of the origin of an earthquake as a single point at some depth below the surface, the wave impulse should move out in all directions, with the wave front at the start exactly spherical; the impulse should first reach the surface at a point vertically above the origin, and the surface area affected should grow from this point in an ever-enlarging circle (Fig. 272). At one time this simple conception of earthquakes was commonly held. The supposed point of origin was called the *focus* or *centrum*, and the point directly above it on the surface the *epicenter* (*e*, Fig. 272). On an ideal isoseismal map (p. 398) the epicenter is at the exact center, for according to the original conception the isoseismal lines should be exact circles, the intensity dying out uniformly away from the point directly above the focus. Usually, however, even where no indication of faulting is visible at the surface, the isoseismals are roughly elliptical, suggesting that the vibrations at the surface started along a line of some length instead of from a point. Not uncommonly, also, in a region visited at frequent intervals by earthquakes the different epicenters are situated along a line; such a line probably represents a long fracture at some distance below the surface.

The terms focus and epicenter still have value in the study of earthquakes, although it should be understood that they are not points. The necessity for modifying the older views is evident when we consider the great length of the fracture on which slipping occurred to cause the San Francisco earthquake. Initial vibrations started along the entire surface of this fault, over a length of at least 270 miles and reaching from the surface to an unknown depth. Isoseismal belts for such an earthquake are extremely elongate.

The general position of the epicenter of a strong earthquake is indicated also by the directions in which objects on the ground are displaced by the shock. At the epicenter the force acts directly upward; in other positions the vibrations emerge obliquely and tend to throw objects away from the epicenter, although monuments and similar objects, from having their bases thrust outward, commonly fall *toward* the center of the disturbance. Therefore the positions of pillars and chimneys that have been thrown down by an earthquake furnish important information.

Seismographs and Seismograms. *Seismographs*[1] are precise and sensitive instruments that record all phases of an earthquake, even at a great distance, with the exact time of each phase. The general principle used in constructing these instruments is most clearly illustrated by one of the older types. A heavy mass of metal suspended like a pendulum has considerable inertia and so tends to remain at rest while the bedrock beneath it vibrates in an earthquake. The suspension of the weight is illustrated in Fig. 273. A concrete base for the instrument is sunk into firm bedrock, and no loose mantle is allowed to come in contact with the base; thus the instrument is kept free from local disturbances, such as jarring due to traffic, which might obscure essential parts of an earthquake record. An upright metal post (P) is bolted firmly to the concrete base, and the heavy weight (*W*), mounted on a rigid arm (*L*), is supported only by a freely moving socket joint (*S*) and a flexible wire (*F*). When elastic vibrations are transmitted through bedrock and the concrete base, the post is set in motion, but the weight remains essentially motionless and so serves as a "steady point."

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[1] Several modern types of seismographs are pictured and described in Reading References 4 and 6 (p. 412). In some of these the relative motion between the weight and its support is utilized to move a small coil of wire in a magnetic field. The small electric current thus generated is much amplified by a vacuum-tube amplifier, then conducted to an oscillograph, where it actuates a galvanometer. Thus the actual vibrations in the ground can be recorded in greatly magnified form.

The recording apparatus is a drum (*R*), which is made to rotate slowly by a clockwork mechanism. This apparatus also is bolted to the concrete base and vibrates with it during an earthquake. These vibrations are graphically recorded, with the aid of the "steady point." In early models a stylus attached to the weight *W* had its point in contact with smoked paper covering the drum. The shaft of the drum is equipped with a worm gear, which shifts the drum longitudinally as it rotates. When the bedrock was undisturbed, the point of the stylus in the old-style seismograph traced a continuous plain line on the smoked paper, as represented at the left in the inset, Fig. 273. When earthquake waves caused the drum to vibrate, however, the stylus recorded the motion as a zigzag line, as shown on the inset. Such a graphic record of earthquake vibrations is called a *seismogram* (Fig. 274).

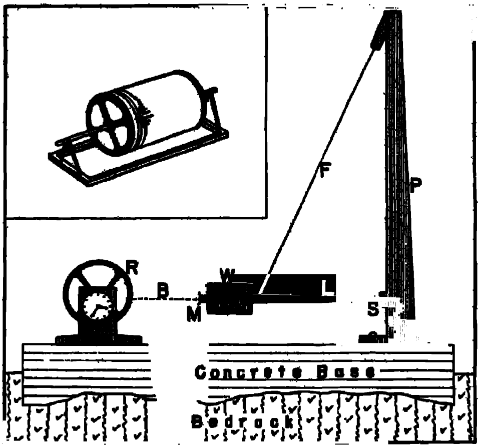

FIG. 273. Generalized representation, in profile, of a seismograph of one type. The chief parts, indicated by letters, are explained in the text. The recording apparatus (*R*) should be much farther from the mirror (*M*) than shown; for this reason the concrete base and the bedrock beneath are represented as discontinuous. Inset at upper left shows in perspective the drum, *R*, with record of an earthquake. (On sensitized paper, the record would not be visible until the paper was developed.)

In a later development of the recording apparatus, still in common use, a beam of light (*B*, Fig. 273) is reflected from a mirror (*M*) attached to the steady point, and impinges on photographic paper covering the rotating drum. The resulting lines are revealed only after the sensitized paper has been developed. Vibrations of a distant earthquake are of course small, and therefore the recording device on most seismographs is constructed to give considerable magnification of the actual amplitudes. All instruments are equipped to mark time intervals on the moving paper, so that each important phase of an earthquake is timed closely (Fig. 275).

Every complete seismograph has three independent components, each making its own record. Part of the vibration in bedrock is in the vertical plane, and one seismographic unit is especially designed to record this vertical motion. In addition, two units for recording horizontal components are mounted at right angles to each other, usually along north-south and east-west lines. The necessity for this is made clear by a simple analysis. Since vibrations are parallel to the direction in which earthquake waves advance, a unit of a seismograph that has the recording device and the "steady point" along a north-south line will give no record of an earthquake whose center is directly north or south of the instrument. Vibrations from such an earthquake will receive maximum recording by a unit mounted along an east-west line. With two units at right angles to each other, a station will give some record of the horizontal motion, whatever the direction of the epicenter from the station.

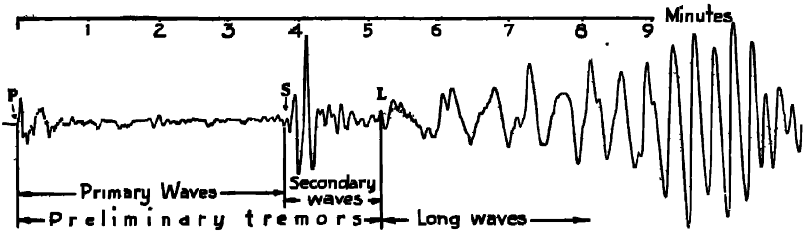

FIG. 274 Record made at Pulkovo, Russia, of an earthquake in Asia Minor. *P* is the beginning of the primary waves; *S*, of the secondary waves; *L*, of the long waves, which continue to the right side of the diagram. The time elapsed between *P* and *S* was 3 minutes and 43 seconds, corresponding to a distance of 1400 miles between the station and the epicentral area.

Modern seismographs are remarkably sensitive and precise, and some of them are made small and light, to permit easy and rapid transportation from one station to another. Portable seismographs are in common use for recording "artificial earthquakes" created by dynamite explosions, to determine underground conditions favorable for accumulations of petroleum and natural gas (p. 523).

The recording of earthquake tremors thousands of miles from the epicenter testifies not only to the sensitiveness of the modern seismograph but also to the high elasticity of the Earth. Seismographs are now widely distributed on all continents, and every major earthquake is recorded in practically every civilized country in the world.

Interpretation of Seismograms. The study of seismograms of distant earthquakes led to the discovery that the main shock is preceded by smaller rapid vibrations known as the *preliminary tremors.* Thus a normal seismogram has the characters seen in Fig. 274. These preliminary tremors represent elastic impulses that come *through the Earth,* in the general direction of a chord from the seat of disturbance to the recording station; whereas the later large vibrations represent those waves that have traveled by a longer route around the circumference. The first preliminary tremors (*P* to *S*, Fig. 274) are caused by a compressional or longitudinal wave (commonly known as the *primary*), which travels several miles per second; the other preliminary set (*S* to *L*, Fig. 274) represents a transverse wave motion (the *secondary* wave), which travels at about half the speed of the primary. Therefore the time interval between the two sets of preliminary tremors is proportional to the distance traversed, and from this information the distance between the seat of the shock and the seismograph can be calculated accurately. The calculation uses the same method required by a problem found in elementary textbooks of algebra, stated as follows: Train *A*, traveling at average speed 50 miles an hour, and train *B* at 30 miles an hour, leave a given station at the same time. *A* arrives at another station three hours earlier than *B*; what is the distance between

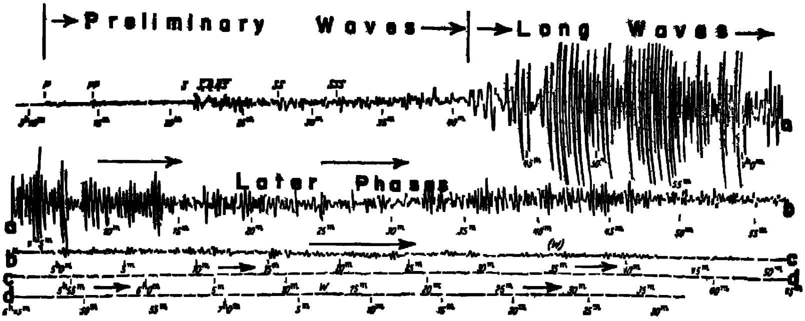

FIG. 275. Seismogram, made by an instrument at Gottingen, Germany, of the Tokyo earthquake, Japan, Sept. 1, 1923, 5700 miles (9200 km.) distant. The first preliminary wave reached the station shortly after 3 A.M., and the record ran continuously for more than 4 hours. (The five segments in the figure fit together as a continuous graph, as indicated by corresponding letters at ends of segments.) Note complexity of preliminary waves and of the impulses following the long waves.

the two stations? In the seismic problem the *P* wave is analogous to train *A*, the *S* wave to train *B*.

If the distances of an epicenter from at least three separate stations are computed and circles are drawn on a globe with these distances as radii, the circles intersect in the epicentral area (Fig. 276). This method of locating the origin of an earthquake invokes a principle that is learned in elementary geometry.

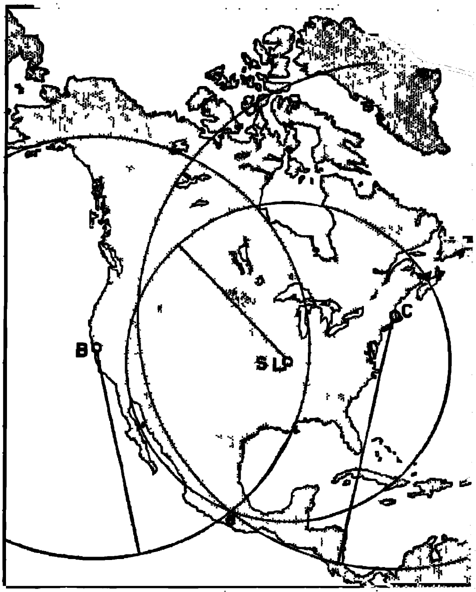

FIG. 276 Method of locating the epicenter of an earthquake from records at three stations. At Berkeley, California (*B*), St. Louis, Missouri (*S L*), and Cambridge, Massachusetts (*C*), the distance is found from the relation of *P* and *S* on the seismogram (Fig 274) Using the distances as radii, circles drawn about the stations intersect in southern Mexico, at the epicenter (*e*).

Most seismograms record a succession of vibrations following the first long waves. Some of these later phases represent recurrent aftershocks, and others are due to complex reflected and refracted wave motions in the Earth (Fig. 275). Every record received at a distant station shows also compound instead of simple *P* and *S* waves (Fig. 275), indicating that refraction of these elastic impulses occurs at the boundaries of definite layers or zones in the Earth that differ in elastic properties (Fig. 285, p. 440).

By careful study of records from several stations distributed around an epicenter, it is possible to calculate the depth at which a shock originated. A large majority of earthquakes that have been studied are in the *shallow-focus* class, having their origin less than 30 miles (50 kilometers) below the Earth's surface. In sharp contrast to these, a considerable number of earthquakes are in the *deep-focus* class, originating at depths as great as 400 miles (700 kilometers). A deep-focus earthquake is felt with nearly uniform intensity in an area hundreds of miles across. Seismograms from such earthquakes are distinctive, since they have practically no record of long waves. Deep-focus earthquakes were first recognized within recent years, and their cause is still a subject of speculation among seismologists and geologists.

With the improvement of seismographs and the wide distribution of stations, seismology has begun to furnish important information about the inaccessible interior of the Earth. This aspect of seismology is discussed in Chapter 18.

Possibility of Predicting Earthquakes. Will the science of seismology be so perfected that eventually we shall be able to predict accurately the occurrence of earthquakes in particular areas? This possibility has been suggested, but probably there will always be too many unknown and variable factors to permit this kind of forecasting with any certainty. The problem has been studied seriously for the vicinity of San Francisco, California. Surveys made during the last half of the nineteenth century, and repeated soon after the earthquake of 1906, suggest the rate at which the rocks near the fault yielded by bending, before elastic rebound on the fault brought relief (Fig. 266). Possibly strain by bending is once more building up, to be relieved by a future jarring displacement. Can the date of this event be fixed?

Any hope of solving such a problem must depend on exact surveys, made at frequent intervals of time, to determine as accurately as possible the rate at which points in the region of the fault are being shifted with the growing strain. Aside from the large expense of making such measurements, the problem involves many difficulties. Although we know the amount of displacement that occurred in 1906 along various parts of the fault, the few measurements made before that date give scant information about the exact positions occupied by critical points outside the fault zone before slipping took place; therefore we do not yet have any dependable standard by which to decide how much deformation of the region is necessary before slipping must occur. We do not know the form of the fault surface underground; if it is irregular, possibly the movement of 1906 caused changes that will make necessary either more or less stress to cause additional slipping. There is no assurance that the forces causing the deformation now act in exactly the same direction, or with the same intensity as before 1906. Moreover, there are other active faults in the same general region, and possibly slipping along one or more of these will ease the strain near the San Andreas fault sufficiently to postpone movement on it for considerable time.

After analyzing all the information available, one geologist suggested that another displacement along the San Andreas fault in central California may occur near the middle of the present century. However, he emphasizes the elements of uncertainty, which may either double or cut in half the average time interval between successive movements. Even such a conditional forecast as this has value because, as he states, "The best protection against the danger of earthquakes is not the knowledge of the particular dates upon which they will occur, but the realization that they may occur at any time, and that foundations and structures should be built sufficiently strong to withstand their shocks."

In many other regions the problem of earthquake prediction is even more difficult than in California. The crustal movements responsible for the Tokyo earthquake of 1923 occurred chiefly under the sea, where no careful check by surveys is possible. Hope of forecasting shocks in such areas must depend on earthquake records over a long period of years, and perhaps on the recognition of minor warning tremors that may herald a major displacement on a fault.

PRECAUTIONS AGAINST EARTHQUAKES

Man can not control the forces in the Earth's crust, but within limits he can protect himself against their destructive effects. Collapse of numerous school buildings in southern California during the Long Beach earthquake of 1933 aroused a widespread demand for earthquake-proof construction of public buildings in that region. Well-constructed buildings with steel frames stood the test remarkably well in Tokyo and in San Francisco. Masonry walls without reinforcement are badly damaged by severe shaking. Tough, resilient materials that recover from strain without breaking stand most successfully, and therefore frame dwellings usually survive a severe earthquake with less damage than those made of brick or stone. It is important, however, to have all buildings as nearly fireproof as possible, since fires from broken electrical connections and from other sources are an incidental effect of every large earthquake. Commonly the water mains are broken when a severe shock affects a large city, and the uncontrolled fires do vastly more damage than the earthquake itself. City engineers in seismic areas are giving special attention to safeguarding water systems. The problem is particularly difficult wherever it is necessary for the mains to cross active faults.

GEOLOGIC EFFECTS OF EARTHQUAKES

Although earthquakes are impressive in their display of power, they play a minor geologic role in comparison with running water and other agents whose work is continuous but as a rule not spectacular. One of the most conspicuous effects of earthquakes is the starting of landslides in mountains or hilly regions. Several large slides of this kind were formed near Tokyo in 1923; and in 1920 the great earthquake in Kansu, China, set in motion enormous quantities of loess, which buried villages and caused important changes in surface drainage. Thus earthquakes speed the process of mass-wasting (p. 60).

Earthquake waves cause alternate compression and tension in the rock or mantle through which they pass. The resulting disturbance causes changes in underground drainage; some old springs stop flowing, others have their discharges increased, and new springs come into existence. Tensional impulses of the earthquake waves open deep fissures in soil or other mantle, and the succeeding impulses of compression force large quantities of water to the surface at some localities. Commonly the flow is concentrated at favorable points along a fissure, with the result that rows of small craters are formed by the sand and silt brought up by the water. Saturated sand in layers far below the surface is forced up to fill crevices, and the resulting "sand dikes" remain as a record of the disturbance. Small dikes of sandstone cutting across shale in old sedimentary rocks may have been formed by ancient earthquakes, although no doubt the pressure responsible for such features originates in various other ways also.

The most important geologic changes at the time of an earthquake are the vertical or horizontal movements of crustal blocks; but it must be emphasized again that such movements are the cause and not the effect of the shocks.

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Chapters I and II give an excellent discussion of earthquakes and their cause. The style is vigorous and stimulating. Numerous excellent illustrations.

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11. The Japanese Earthquake of 1923; by Charles Davison. 124 pages. Thomas Murby and Co., London, 1931.

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Detailed and highly interesting accounts of sixteen great earthquakes, ranging in date from 1755 to 1931.

CHAPTER 17

METAMORPHISM

Metamorphism the Result of Adjustment to Environment. The Earth's outer shell in many places is made up of remarkable rocks that are obviously neither sedimentary nor igneous. These rocks, known as *metamorphic*, comprise the third of the three major groups into which all rocks are divided. Metamorphic means *transformed* and tells us that the rocks so called have been made by the transformation of older, pre-existing rocks. This transformation has developed new structures, new textures, or new minerals, or all of these. The resulting metamorphic rock may consequently have been so thoroughly made over and have become so changed in appearance that it would not be suspected to be an older rock in a new guise. In places, however, metamorphic rocks grade by imperceptible stages into rocks known to be of sedimentary or igneous origin. Such gradations furnish the clue to understanding how the metamorphic rocks came to be.

Among the surprising discoveries made by the microscope is the fact that many rocks are adjusted to their environment. The minerals that occur together in a given rock are more or less thoroughly adjusted to the environment in which the rock was formed. If they are in complete adjustment, they are in stable equilibrium under the conditions prevailing in that environment. Under changed conditions they cease to be stable. They therefore tend to adjust themselves to the new environment, and new minerals are formed.

Metamorphism accordingly is the result of the tendency of rocks to adjust themselves to their environment. The factors determining equilibrium between minerals in rocks are temperature, pressure, and composition. Change in any one of these factors disturbs equilibrium; and to restore equilibrium internal changes take place in the rocks. Take for example a sedimentary rock. It was formed at the temperature and pressure prevailing at or near the Earth's surface. If later it becomes involved in crustal folding, it may become deeply depressed in the crust, miles below the surface. During folding it was subjected to strong confining pressure and to powerful shearing stresses; and, when it reaches its final position deep in the crust, it is in a greatly different environment, where temperature and pressure are far higher. Many of its constituent minerals are unstable under the changed conditions and react with others present to form new minerals that are stable under the new conditions. As a result a new rock is formed.

Rocks in which the minerals are in equilibrium with one another are termed *equilibrium rocks.* Because silicate minerals react with one another extremely sluggishly, new equilibrium has not always been attained under the changed conditions. Such rocks in which equilibrium is not complete are *disequilibrium rocks.* The disequilibrium rocks contain relicts of their former state and are therefore more interesting than the equilibrium rocks. Thus they give us evidence not only of what they were before metamorphism, but also of the direction in which they have been changed.

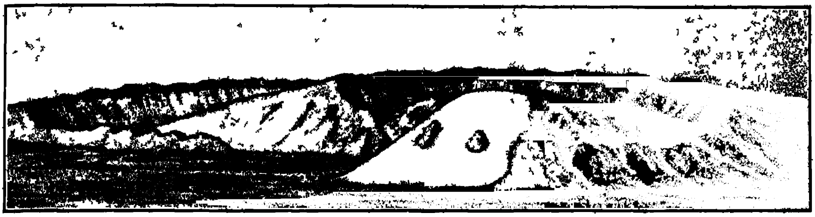
Metamorphic rocks have been formed not only from sedimentary and igneous rocks, as long recognized, but also from older metamorphic rocks.

PHYSICAL-CHEMICAL FACTORS OF METAMORPHISM

Temperature. Temperature is one of the most powerful factors, not the most powerful factor, in causing metamorphism. Rise of temperature speeds up chemical reactions; but the much more important result of rise of temperature is that some reactions take place at the higher temperatures that do not take place at lower temperatures. For every aggregate of minerals a critical temperature must be reached before chemical reaction can begin to take place between the minerals, and only then can new minerals begin to form. The theoretical explanation is that the atoms that make up the minerals must attain a certain amplitude of vibration before they can change places with others. This principle has been established experimentally in the laboratory and is of the highest importance in understanding metamorphic phenomena. As a concrete example: one of the commonest metamorphic reactions is that in which white mica reacts with chlorite to form the new mineral biotite; but unless the mica and chlorite are subjected to a temperature that exceeds the critical temperature at which the amplitude of thermal vibration has become so large that reaction between the minerals begins, they will coexist side by side throughout geologic time.

Temperature is also highly effective in increasing the plasticity of minerals, thereby increasing the deformability of the rocks in which they occur.

Pressure. Pressure is another important factor in metamorphism. Its effects differ greatly whether it is *uniform* (all-sided, or confining pressure) or is *nonuniform*. A uniform pressure is one in which the applied force acts equally on all sides of a body. A nonuniform pressure is one in which the applied force is greater or less in one direction than the forces applied in the other directions; as a result of this set-up there is a differential pressure. As a matter of fact, it is differential pressure developed at high confining pressure that produces most of the notable structures in metamorphic rocks. Most of the effects ascribed to pressure in geology are those produced by nonuniform pressure.



Modified from G. M. Lees.

FIG. 277. Salt "glacier," Iran. The salt, which is being extruded from a plug that has pierced through an anticline to the surface, flows downhill, like a glacier of ice.

The behavior of ice illustrates some of the functions of pressure in the processes of metamorphism. Ice under ordinary conditions is easily broken upon slight impact: it is a highly brittle solid. But ice confined in a steel jacket, which is fitted with a piston, can be made to flow through an orifice in the jacket by subjecting the piston to heavy pressure. The ice within the steel jacket, because it is under a high confining pressure, has been caused to flow through the orifice by the differential pressure. Ice thus confined and subjected to a differential pressure is no longer brittle but has become plastic, which means that it has become capable of continuous and permanent change of shape without breakage. The counterpart in nature of such plastic ice is the flowing ice beneath the carapace of brittle ice that envelopes the upper and outer portions of a glacier. Similarly rock salt, brittle at ordinary pressure, becomes plastic under high confining pressure and if subjected to differential pressure can be made to flow, as experimentally proved. This property of rock salt to flow plastically has been confirmed by the astounding discovery of salt "glaciers" in the rainless regions of Persia. Salt plugs (p. 521) have punched their way to the Earth's surface, and, being still under pressure, the salt is squeezed out and flows away like a glacier (Fig. 277). The longest of the salt glaciers is 2 miles long.

Limestone and marble also have been made to flow plastically in the laboratory, and Nature has forced them to do so in an immensely larger way, in the Earth's crust. Brittle and plastic are relative terms, and probably all minerals and rocks if under strong enough confining pressure and if subjected to powerful differential pressure slowly applied, would flow by plastic deformation. The plastic deformability of minerals is greatly increased by temperature, probably doubling for every 10-degree rise in temperature.

The forced flowage of minerals and rocks develops enormous shearing stress within them. As has long been indicated by the evidence from metamorphic rocks, such shearing stress is highly effective in causing mineral transformations. This remarkable effect of shearing stress has been experimentally proved by Bridgman. By subjecting substances to high confining pressure and giving them a sudden twist, thereby developing a shearing stress, he was able to cause a large number of chemical reactions, without rise of temperature. Shearing stress acted as a sort of catalyst, promoting reactions that would not otherwise have taken place. Although no geologic reactions were reproduced, it is highly probable that when experiments are undertaken on minerals at higher temperatures, or the shearing stress is developed more slowly, or both, they will succeed in forming new minerals.

In many rocks the effect of forced flowage is to crush, mash, or pulverize the component minerals, as is so astoundingly shown in mylonites, described on page 423. In these rocks the minerals were unable to flow plastically, possibly because of insufficient confining pressure, too rapid application of the deforming forces, or too low temperature, or of inadequacy of all these factors Rocks thus deformed are said to have been mechanically metamorphosed.

Some rocks in which the metamorphism was chiefly mechanical nevertheless contain some newly formed minerals, such as white mica (sericite), which can be seen under the microscope to have developed, probably as constructive effects of shearing stress. From such incipient stages, we can find a complete gradation to rocks that consist entirely of newly formed minerals. In the completely reorganized rocks all evidence of the deformation through which they went earlier has been obliterated. Transitions of the kind just mentioned support the idea, that many if not most metamorphic rocks have resulted from shearing stress developed by forced flowage. Since evidence of internal movement within the rocks can often be found, metamorphism of this kind is called *kinetic metamorphism*. As the operation of powerful forces is indicated, it is also called *dynamic* metamorphism, and because high temperature doubtless cooperated in causing the thorough reconstitution of the rocks it is also called *dynamothermal* metamorphism.

If new minerals are being formed in a rock that is being metamorphosed, high confining pressure favors the development of minerals that contain the most matter in the least space; in other words, heavy or dense minerals. High confining pressure thus tends to favor the formation of space-saving minerals. Garnet is a notable example of such a space-saving mineral. Metamorphic rocks formed under great pressure are therefore characterized by their high specific gravity.

Composition. Composition is the third factor in metamorphism. If the adjustment by metamorphism is complete, rocks of identical composition, regardless of their original condition or origin, yield identical metamorphic products. For example, granite, rhyolite, and arkose, rocks that are identical in composition but are greatly different in appearance and origin, after thorough metamorphism become entirely similar.

In some kinds of metamorphism as described on page 422, various chemical constituents added from outside sources change the original composition of the rock.

Inasmuch as metamorphic rocks have been formed from igneous, sedimentary, and pre-existing metamorphic rocks, they show extraordinary diversity.

Temperature, pressure, and composition are the three physical-chemical factors that determine the nature of the resulting metamorphic rock. It is the geologist's job to determine what geologic processes acting in the Earth's crust have supplied the necessary temperature and pressure to produce metamorphism.

GEOLOGIC FACTORS OF METAMORPHISM

Thermal Metamorphism. Transformation effected in rocks mainly by high temperature under static conditions is called thermal metamorphism. The high temperature can be produced in two ways: by intrusion of igneous masses, and by deep burial in the Earth's crust.

Contact, or Igneous, Metamorphism. The rocks that border an intrusive igneous mass have generally been notably changed from their normal condition at some distance from the igneous mass. They have been changed, or metamorphosed, chiefly by the high temperature to which they were subjected by the magma and the hot gases that issued from the magma when it solidified. The changes are of course greatest adjacent to the igneous mass and fade away outward from the contact;

hence metamorphism of this kind is termed *contact metamorphism.* To signalize the agent that caused the metamorphism, the term *igneous metamorphism* is often used. An obvious example of contact metamorphism, probably the most impressive because coming more nearly within the ken of ordinary experience, is a coal bed that has been converted into a mass of coke by a sill injected parallel with the coal bed just above or below it. Beds of natural coke produced in this way are found in Colorado and in other regions where coal-bearing strata have been invaded by igneous bodies.

As an illustration of metamorphic phenomena, contact metamorphism is particularly enlightening. In its simpler manifestations we see the effect of one factor of metamorphism—temperature, uncomplicated by the effects of other factors Moreover, we can often trace the transition from the highly metamorphic product at the contact to the original unaltered rock distant from the contact. We can therefore ascertain which new minerals form at moderate temperatures and which at high temperatures, and thus we obtain a scale for measuring the intensity of metamorphism.

Geothermal Metamorphism. Some rocks appear to have been metamorphosed because they were at one time deeply buried, having become depressed in the crust under a heavy load of overlying rocks. The higher temperature brought about by the deep burial has caused new minerals and textures to develop, and the heavy pressure due to the load favored the production of heavy minerals, such as garnet, which is a space-conserving mineral. As the Earth's own heat is the main factor here in transforming the rocks, the process is called *geothermal metamorphism.*

The best-established example of metamorphism of this kind is furnished by the German potassium-salt deposits. These salts were laid down in an evaporating arm of the sea in Permian time; they were of many kinds and of complex compositions, but they were stable under the conditions of moderate temperature that prevailed when the sea was evaporating. Later the basin in which they were deposited slowly subsided, and they became covered by 20,000 feet of sedimentary beds. The temperature of the deeply buried salt beds rose about 200°C. to the temperature determined by their depth in the crust. The original assemblage of salts was no longer in adjustment with its environment; consequently the mineral composition of the salts was rearranged, and many new minerals were formed.

Coal is another substance peculiarly sensitive to geothermal metamorphism. Its metamorphic rank (p. 313) increases with depth; for every hundred feet of depth its fixed carbon content, which serves to measure its rank, increases about 0.6 per cent, though this figure differs somewhat in different coal districts.

Silicate rocks are far less sensitive to geothermal metamorphism than coal and the minerals of the marine salt deposits. Nevertheless, certain formations are believed to have acquired their metamorphic state by geothermal metamorphism. In places the bottoms of geosynclines have become so deeply depressed in the crust by sedimentary loading and subsequent crowding together of folded strata (p. 426) that the rocks have been raised to high temperature and have become metamorphosed.

Kinetic Metamorphism. Transformation effected by nonuniform pressure is called kinetic metamorphism. It may be, as already explained, purely mechanical, causing the crushing of the constituent minerals of the rocks without being accompanied by the formation of new minerals. More generally, the rock deformation is accompanied by or followed by the growth of new minerals, culminating in complete reconstitution of the affected rock.

Injection Metamorphism. The changes effected in the rocks of a contact-metamorphic zone surrounding a batholith emplaced under mountain-building pressures are comprised in the term *injection metamorphism.*

CONTACT METAMORPHISM

Every intrusive igneous body—dike, sill, volcanic neck, laccolith, stock, and batholith—is bordered by a zone of baked and hardened rocks generally of considerable width. Extrusive rocks, such as lavas, however, are able at most to bake only slightly the soils or rocks over which they flow.

The altered state of the rocks surrounding an intrusive igneous mass is the result of the growth of minerals that require high temperatures to form them. The zone of altered rocks is called the contact-metamorphic zone. The intensity of the changes diminishes and fades out at some distance from the contact with the intrusive mass; hence the zone thus altered is styled the *contact-metamorphic aureole.* Aureole, meaning "halo," is an apt descriptive term in this connection.

Dimensions of the Contact-Metamorphic Zone. The thickness of the contact-metamorphic zone depends on several factors: on the size of the igneous mass, on how susceptible to metamorphism the rocks bordering the igneous mass are, and on other conditions. The thickest zones surround stocks and batholiths. They may be a mile or more thick but are generally much thinner. Adjacent to small intrusive masses, such as thin dikes and sills, the zone ranges from a few inches to a few feet.

The width of the contact-metamorphic zone around an igneous mass may vary, for it is controlled by the form of the igneous mass and by the attitude of the surrounding rocks. Thus in Fig. 278, which shows an igneous mass (a stock) intrusive into a series of horizontal beds, the contact-metamorphic zone as exposed at the Earth's surface is narrowest on the side of the stock where the contact plunges steeply downward, and it is widest where the contact surface has the smallest inclination. The variable width of the contact-metamorphic zone on the Earth's surface is thus purely a geometric effect, for the thickness of the zone, which is measured perpendicularly to the contact surface, is everywhere the same. On the other hand, the attitude of the beds that surround the intrusive mass—whether they dip toward it or away from it—affects the actual thickness of the aureole (Fig. 198, p. 290), because heat is transmitted more readily parallel to the beds than across them, both by conduction and by emanations traveling through them.

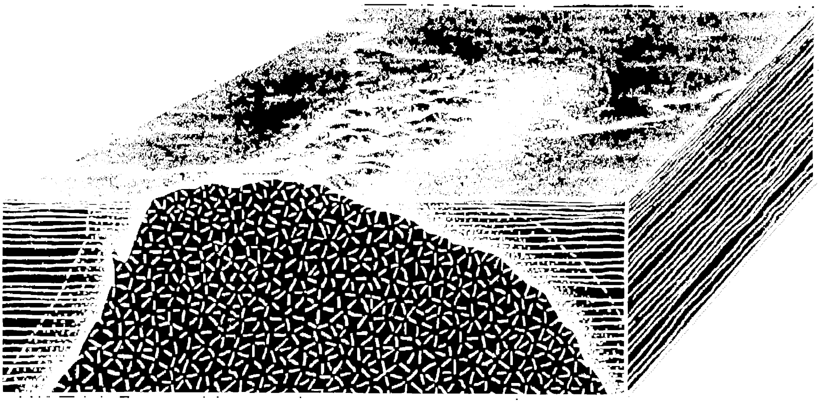


Fig 278. Stock, intrusive into horizontal strata and partly bared by erosion The width of the contact-metamorphic zone (stippled) as measured on the Earth's surface is seen to depend on the dip of the contact surface the zone is widest at the right, where the dip of the contact surface is least steep. Scale is roughly 3 miles to the inch.

Contact Metamorphism of Rocks of Different Kinds. The effect exerted by an igneous mass on the surrounding rocks depends greatly on the nature of those rocks. Pure sandstone is but slightly affected, though near the contact it is changed into quartzite—a compact rock so firmly cemented that it fractures across the grains, instead of around them as in the unaltered sandstone. Shale, however, shows notable effects: it is greatly hardened, and near the contact is converted into a fine-grained rock known as *hornfels*, which to the unaided eye strongly resembles a black flint or a fine-grained igneous rock, such as basalt. Examined microscopically, the hornfels is found to be a mass of newly formed minerals. Some hornfelses are dotted with crystals of *andalusite*, a half inch or more in length, so that they resemble porphyries. To form the andalusite crystals there has been a notable accretion of certain atoms around centers of crystallization, and the growth of these crystals during metamorphism testifies to the remarkable migratory power imparted to the atoms by the rise in temperature.

Pure limestone is strikingly changed through considerable distances from the contact. Dull, somber-colored aphanitic limestone is transformed to a sparkling snow-white marble visibly crystalline to the unaided eye. The heat of the igneous mass has stimulated the minute calcite grains of the limestone to grow in size, has driven off the carbonaceous pigment, and has thus produced a marble of dazzling whiteness.

Limestones that contained sandy or clayey impurities show highly interesting changes. They too were turned into marbles, but these marbles carry in addition many newly formed minerals. The minerals were formed by chemical reactions that took place between the impurities and the calcite (calcium carbonate) of the limestone, reactions that can occur only at high temperatures. Thus when a limestone containing silica as an impurity is heated above 500°C., carbon dioxide is driven off, just as in "burning" limestone, and the lime and silica unite to form the metamorphic mineral *wollastonite*, a calcium silicate. If the limestone contains also dolomite (calcium-magnesium carbonate) as an impurity, then *pyroxene*, which is a silicate containing magnesium in addition to calcium, is formed instead of wollastonite.

The pyroxene thus produced occurs in the inner, hotter portion of the contact-metamorphic zone, whereas a mineral of somewhat similar composition but of different crystal form, a white amphibole (tremolite), is formed in the outer, cooler portion of the zone. Consequently, these two minerals, pyroxene and amphibole, are useful as "geologic thermometers" for measuring the temperatures at which metamorphism is effected. The pyroxene, requiring for its formation a higher temperature than the amphibole, is said to be a *higher-rank* metamorphic mineral than the amphibole.

If the limestone contains other impurities such as clay (which furnishes alumina and iron oxides), garnet and other new minerals are formed. If the impurities are sufficiently abundant, the resulting product is an aphanitic white or cream-colored rock—a *calcic hornfels.*

Coarse-grained igneous rocks, being the products of magmatic solidification and having therefore been already at high temperatures, are generally but little affected by later intrusions. A granite intruded by a younger granite remains unaffected. However, volcanic rocks, such as basalts, if invaded by plutonic rocks, as has occurred in many regions, become drastically metamorphosed. The conditions under which volcanic rocks form differ so markedly from those under which plutonic rocks form that when volcanic rocks are invaded by plutonic rocks far-reaching changes are necessary to adjust them to the new environment created by the intrusion.

The effects produced in a contact-metamorphic zone are as a rule due mainly to the heat given off by the igneous mass. The rocks that surround the igneous mass become highly heated, and new minerals grow in them by recombination of the elements already present. The resulting metamorphic rocks have therefore the same chemical composition as the rocks from which they were derived. A marble, for example, has the same chemical composition as the limestone from which it was formed. Contact metamorphism of this kind, in which the rocks remain unchanged in chemical composition, is by far the most common type of alteration caused by intrusive igneous masses. It produces a contact-metamorphic zone that encircles a stock or batholith as a continuous belt. As the zones may be a mile or more thick, large volumes of metamorphic rock are formed. Because contact metamorphism of this kind is widely prevalent, it is termed *normal contact metamorphism.*

Addition of New Substances to the Contact Zone. An igneous mass while solidifying may give off gases that carry iron, silicon, boron, and other elements. Changes that differ vastly from those brought about by normal contact metamorphism are then produced, especially in limestone, because limestone reacts readily with the hot gases traversing it. Consequently, many new minerals are formed in the limestone, as a rule beautifully crystallized. The resulting contact-metamorphic rock differs greatly in composition from the original limestone: it has had added to it a large amount of substances brought by the gases that streamed from the magma. These gases after their release from the magma generally travel by way of fissures through the surrounding rocks, and the effects they produce are localized along these channel ways. Consequently, metamorphism of this kind does not form a continuous aureole surrounding the stocks and batholiths, such as is characteristic of normal contact metamorphism. As gases set free from the magma are the means by which the substances are transferred from the magma to the surrounding limestone and cause the reactions in the limestone, contact metamorphism of this kind is termed *pneumatolytic* (new-măt-o-lit'ik).

If the magmatic gases contain iron, copper, or tungsten in notable quantity, valuable ore deposits are formed in the limestones as by-products of metamorphism of this kind. For example, most of our domestic supply of the exceedingly valuable steel-alloy element tungsten comes from contact-metamorphic deposits formed near granite contacts. In mining ore deposits that were formed in this way, an understanding of the laws of contact metamorphism is of prime importance.

KINETIC METAMORPHISM

Metamorphism has frequently accompanied deformation of the Earth's crust. As a result of folding, slipping takes place between adjacent strata, and frictional drag develops. Under some circumstances, this frictional drag, by producing intra-mineral and inter-mineral movement within the rocks affected causes metamorphism. The qualification "under some circumstances" is necessary, because many strata have been acutely folded but were not thereby metamorphosed. It takes more than folding to bring about metamorphism in folded strata.

Rocks adjacent to certain great thrust faults, where the movement has not been concentrated along a single surface but was distributed along many closely spaced parallel surfaces (=zones of distributive movement), have been greatly metamorphosed. Metamorphism of this kind is accompanied by tearing, stretching, and mashing of the original elements in the rock, so that, for example, the pebbles of a conglomerate are drawn out into long pencils. As these effects vividly suggest that enormously powerful forces have brought about the metamorphic changes, this metamorphism is called *dynamic metamorphism.* The term *kinetic metamorphism* is, however, preferred because it emphasizes the fact that the rocks so metamorphosed show evidence of internal *movement.*

Mylonite. Where great slices of the crust have been shoved over the underlying rocks, as in the thrust faults described on page 465, the rocks near the fault surfaces are likely to show signs of great mechanical disturbance, with little or no accompanying formation of new minerals. Along some of the great thrust faults the rocks have been so damaged that their original characters are no longer recognizable.

The constituent minerals were mashed, pulverized, and dragged out; as a result a streaky or banded, compact, flinty rock is produced, termed *mylonite* (from the Greek *mylon,* a mill), expressive of the fact that a mylonitized rock has been through the metamorphic mill. The terms "mashing" and "pulverizing" call emphatic attention to the most striking feature of mylonitization, namely that an immense number of minute grains have been produced at the expense of the original grains in the rocks mylonitized. They are somewhat misleading, however, because the mylonites have not lost their coherence or strength.

Rocks ranging from those as weak as coal to those as strong as granite have been reduced to mylonites. In Fig. 279 is shown a zone of mylonite, formed along the base of an overthrust mass of granite and now exposed by erosion; at the time of thrusting and mylonitization the thickness of granite above the fault surface was doubtless much greater than it is now. If the transition from mylonite to undamaged granite were not traceable in the field, the origin of the mylonite would remain enigmatic, so completely does the mylonite differ from the granite from which it is derived.

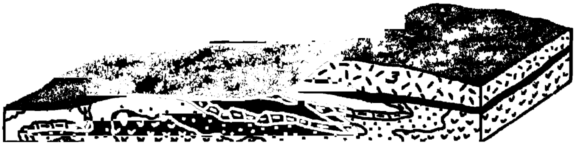


FIG. 279. Zone of mylonite (2) formed by frictional drag as the result of thrusting of the granite mass (3) over the underlying rocks (1).

The nature of the resulting mylonite is of course influenced by the mineral composition of the original rock. If, for example, a granite containing considerable black mica is mylonitized, the mica will be smeared out into an immense number of infinitesimal flakelets, all in parallel arrangement, and the resulting rock resembles a black slate or a phyllite. Phyllite of mylonitic origin is termed *phyllite-mylonite,* or *phyllonite,* for short.

A mylonite is the result of intense mechanical deformation: it is the finest example of a metamorphic rock formed by purely mechanical processes, uncomplicated by the growth of new minerals. However, there are transitions from such purely mechanically deformed rocks to those in which some new minerals have developed, and finally to rocks in which all the minerals have been newly formed and all evidence of the milling through which the rock went has disappeared.

Such transitions support the current theory that many foliates are the results of kinetic metamorphism.

Foliates. In addition to the rocks of contact-metamorphic origin, an immensely larger group of metamorphic rocks has been formed by rock movement generally but not always accompanied by mineral crystallization. Most distinctively, these rocks have a structure parallel to which they tend to split into flakes, leaves, or thin slabs. This structure is the result chiefly of the parallel arrangement of platy or lamellar minerals developed during or after the rock movement. It is termed *foliation* (from *folium*, a leaf), and rocks that have this structure are *foliates*.

The most abundant as well as the most effective foliation-producing minerals are the micas (muscovite and biotite) and chlorite. These minerals normally occur as thin plates and have a perfect cleavage parallel to the platy surface. Hence a rock that contains many plates of these easily cleavable minerals, all of which are in approximately parallel orientation, will readily split parallel to the planes in which they lie.

A foliate in which the foliation is poorly defined is called a *gneiss* (nīce). A foliate in which the foliation is well defined and closely spaced is termed a *schist* (shĭst; from the Latin *schistus*, that which can be split). No sharp boundary divides gneisses and schists, for a complete gradation between them exists in nature.

A well-foliated rock in which the minerals are so fine grained that they are no longer recognizable by the unaided eye is termed a *phyllite*. In *slate*, which is still finer grained than phyllite, the foliation caused by the parallel arrangement of innumerable minute flakes is so perfect that the rock can be split along closely spaced surfaces that are practically smooth planes.

Gneisses and schists are collectively known as *crystalline schists*, a holdover term from the time before the microscope was used to study rocks. It refers to the fact that gneisses and schists are made up of grains visibly crystalline to the unaided eye, in contrast to the phyllites and slates whose constitution was unknown.

Foliates are widely distributed over the Earth, and in some large regions they are the only rocks exposed. All of the interior of Alaska, for example, is underlain by them. They are well shown in the inner gorge of the Grand Canyon of the Colorado. They occur also in the cores of many mountain ranges, where they have become exposed by deep erosion.

REGIONAL METAMORPHISM

Regions of metamorphic rocks, if of uniform metamorphic intensity or if the intensity changes gradually and systematically across the region, are said to show *regional metamorphism.* The term regional metamorphism has the merit of being noncommittal, for it leaves open the question whether the temperature required to cause the metamorphism is the result of igneous influences, such as deep-seated batholithic invasion of the crust, or was produced by geothermal temperature, or by unknown causes.

A hypothesis now much in favor holds that regional metamorphism is one of the effects that accompanies the folding of a geosynclinal mass of sediments. As the result of the folding and crowding together of the strata during a mountain-making revolution, a heavy load, greater than the strength of the crust can endure, is placed upon a portion of the geosyncline, which therefore sinks in response to isostasy. The most heavily loaded portion sinks the deepest. This portion of the geosyncline thus becomes exposed to the heat of the deeper zones of the Earth's crust. Metamorphism therefore sets in. Subsequently the fold mountains born of the geosyncline (p. 463) are attacked by erosion, and eventually the roots of the mountains are brought to the Earth's surface in accordance with the principle of isostasy (p. 18) and become accessible to view. Gneisses and high-rank metamorphic rocks are seen to have developed in the highly folded, deeply subsided portion of the geosyncline; laterally they grade gradually into mica schists, and still farther outward the schists grade into phyllites and slates. Thus the metamorphic tract shows a systematic arrangement in belts of differing metamorphic intensity, decreasing laterally outward from the central belt.

The highly folded rocks in the deeply depressed portion of the geosyncline have generally been invaded by great batholiths of granite. Differences of interpretation have therefore arisen as to how much of the metamorphism was effected by the folding, by the Earth's own heat at great depth, and by the heat of the batholithic magmas. The attack on these problems is part of the advancing front of geologic science.[1]

COMMON VARIETIES OF METAMORPHIC ROCKS

Slate. Slate is a minutely grained foliate, whose most distinctive feature is its capacity to split or cleave into thin parallel sheets. This property, the well-known *slaty cleavage,* reaches its acme in roofing slate, which can be split into extremely thin, smooth parallel sheets, down to one tenth of an inch in the most cleavable slate. Such remarkable cleavage is described as plane-parallel.

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[1] See note on page 436

Slate is a metamorphic rock of the lowest rank of metamorphism. Some indeed—clay slate—is shale on which merely a slaty cleavage has been impressed. Other slate—mica slate—not only has acquired a slaty cleavage, but also has had its mineral composition reorganized by the growth of new minerals, most commonly white mica in ultrafine scales. Thus mica slate represents a stage of progressive metamorphism one step higher than that of clay slate.

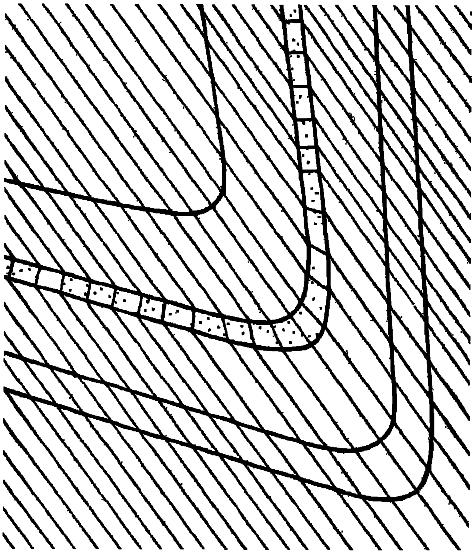
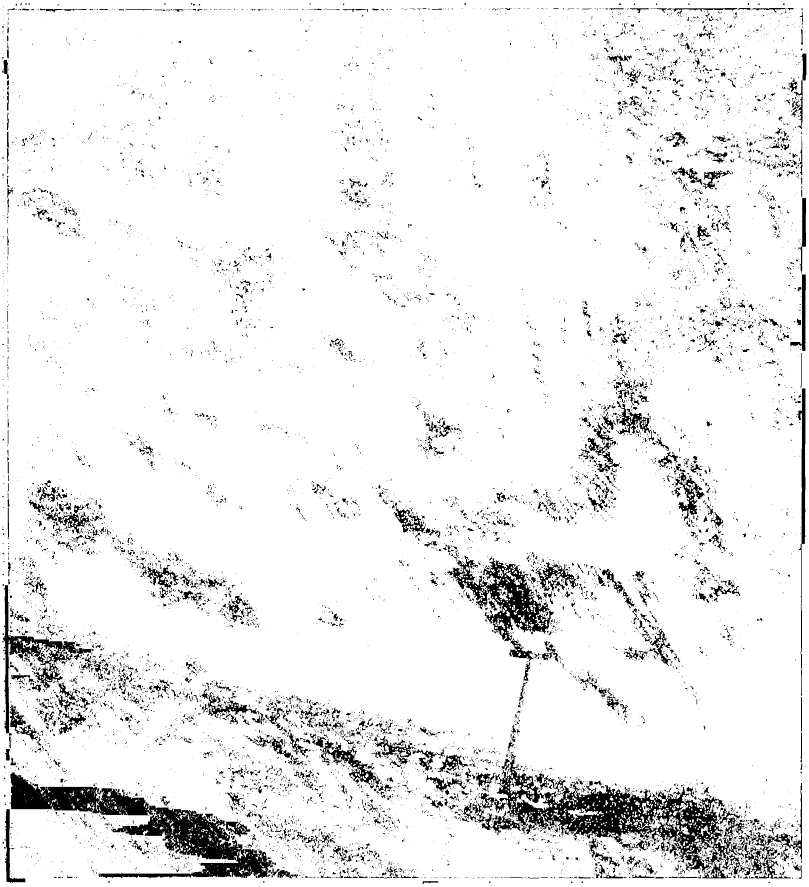


Fig. 280. Slaty cleavage cutting the bedding of folded beds. Refraction of cleavage is shown on entering a harder bed (sandstone, indicated by stippling).

Slaty Cleavage. The cleavage of folded strata that have been converted into slate is parallel to the axial planes of the folds (Figs. 280 and 281). As a result the cleavage cuts across the bedding of the slate in the arches and troughs of the folds and tends to be parallel to the bedding in the limbs of the folds. The closer the folds have been pressed together the more nearly the cleavage is parallel to the bedding in the limbs of the folds.

The trend of the cleavage in a slate district maintains a constant direction. Unless we use keen discrimination, we can easily mistake slaty cleavage for bedding; and, if the cleavage is thus misinterpreted, the inferred geologic structure of the area in which the slate occurs will be in error.



Arthur Keith, U. S. Geological Survey.

Fig. 281. Slaty cleavage cutting across the beds of an overturned syncline. Cleavage is parallel to the axial plane of the syncline, dipping 60° from left to right. Walland, Tennessee.

The cause of the remarkable plane-parallel cleavage of slate is revealed by the microscope: it is a property determined by the fact that slates are largely built up of innumerable closely spaced minute flakes of micaceous minerals that are in parallel arrangement. These flakes, mainly white mica and chlorite, have an excellent cleavage parallel to their flat surfaces; and the cleavage of the slate is therefore in essence a summation of the marked cleavage capacity of these innumerable parallel flakes.

Why these minerals have this strict parallel orientation in slate is another problem, which has not yet been solved. Because most slate occurs in folded strata and the fossils it happened to contain are deformed and flattened in the direction of the cleavage, directed pressure was clearly a factor in developing the cleavage. Shale is the rock most commonly transformed into slate.

Phyllite. Phyllite is a foliate that is intermediate in metamorphic rank between slate and schist. That at the low-rank end of the sequence resembles slate, but its constituent mica is coarser, which gives it a glossy, silky luster. The minerals sandwiched between the mica flakes are also somewhat larger than they are in slate, and because of this coarser grain size the foliation of a phyllite is slightly less closely spaced than in slate and approximates smooth planes less closely.

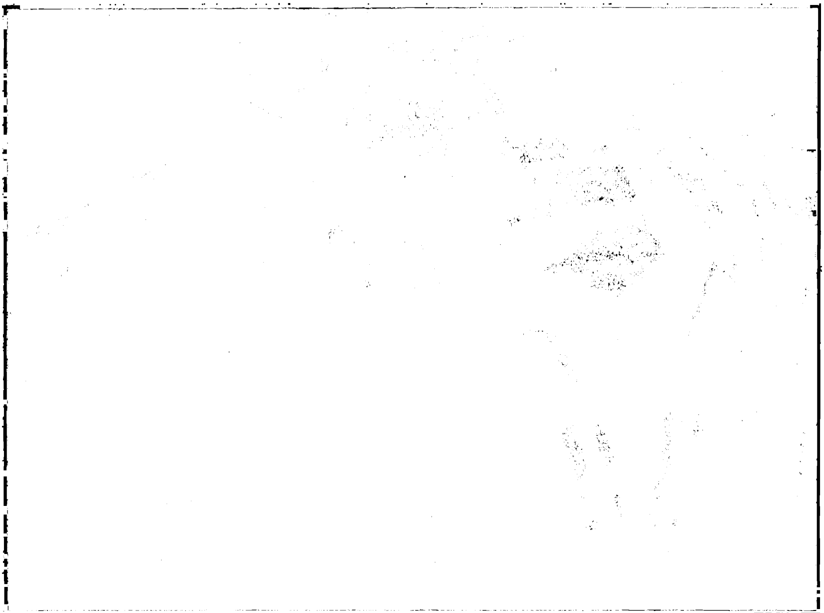
Some phyllites contain sporadic prominent crystals, such as garnet. These minerals are sure indices of the higher metamorphic rank of phyllites and clearly distinguish them from slates. Phyllite that is near the higher-rank end of the series between slate and schist resembles schist but is finer in texture.

Schist. Schist is the most abundant member of the great group of metamorphic rocks called crystalline schists. It is visibly crystalline to the unaided eye. Its foliation is thinly spaced, but it has lost the plane-parallel character of the foliation of slate. The far greater grain size of the minerals that make up the schist causes the foliation surfaces to be uneven and more or less wavy. Foliation of this kind is called *schistosity*.

There are many varieties of schist. They are named according to their most distinctive constituent: thus we have mica schist, hornblende schist, and chlorite schist, to mention only the more prominent kinds.

Mica schist is the most abundant kind of schist. Its essential minerals are quartz and mica, and the numerous flakes of mica give the schist its highly characteristic foliated character. Different varieties of mica occur; the most abundant are silvery white muscovite, which gives the rock a brilliantly spangled appearance, and the black mica, biotite. The micas are irregular flakes, all arranged with their flat sides in parallel orientation. The cleavage planes of the micas are thus oriented in one plane, which is the plane of the schistosity. This parallel arrangement of the micas and their cleavage planes produces the well-defined foliation of mica schist. Some mica schists contain large prominent crystals of garnet, staurolite, and other minerals, which give the schists in which they have grown a texture simulating the porphyritic texture of igneous rocks.

Mica schist is a rock of high metamorphic rank. The original material from which most mica schists were derived was an argillaceous sedimentary rock—a shale of some kind—and the transformation of the dull, amorphous substance of the shale to a brilliantly spangled mica schist is in a way as remarkable as the metamorphosis of a chrysalis into a beautiful butterfly.



P. J. Holmquist.

FIG. 282. Folded gneisses and intercalated layers of marble (brilliantly white). Near Stockholm, Sweden.

Gneiss. Gneiss is an imperfectly foliated rock of granular appearance to the unaided eye. Much gneiss is streaky or banded, because it is made up of alternate layers of differing mineral composition. Some gneiss that was formed by the metamorphism of sedimentary rocks has inherited a remarkably even, regular banding or layering, as shown in Fig. 282.

Gneiss is abundant and includes many varieties formed under very diverse conditions.

Some gneiss was formed under conditions of mechanical metamorphism in which its minerals were mashed or otherwise mechanically deformed. Figure 283 shows a gneiss derived from a granite by kinetic metamorphism that has not gone beyond the stage of mechanical deformation. The granite was subjected to differential heavy pressure; the feldspars were partly crushed, the quartz was "pulverized," and the biotite was shredded and dragged out into long streamers. The result is a granite gneiss of strongly marked characters: it is of streaky, banded appearance and has a rude, imperfect foliation. The ultimate result of more intense deformation of the kind to which this gneiss was subjected is a mylonite, as already described.

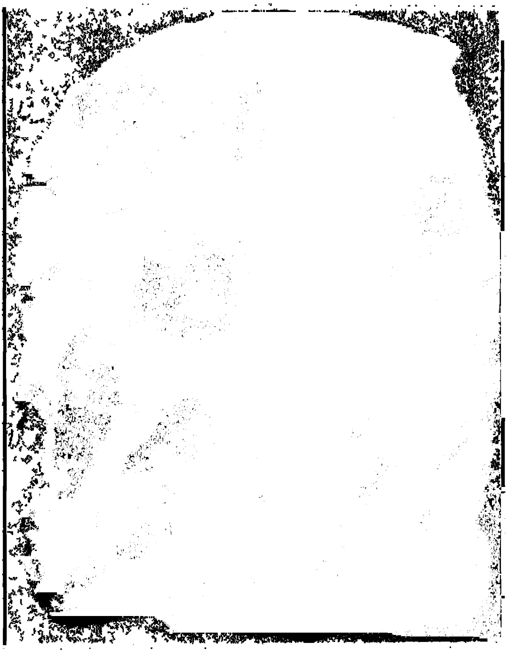


Fig. 283. Granite gneiss, formed by mechanical metamorphism. As result of forced flowage in the solid state a coarse foliation was impressed on the granite; large orthoclase crystals had their corners rubbed off and their shapes changed to ovoid and tadpole-like forms.

Many gneisses, however, are high-rank metamorphic rocks. They are characterized by containing minerals, such as garnet and pyroxene, that were newly formed at high temperatures and pressures.

Some granite gneisses are called *primary gneisses*, because they took on a foliation while they were still only partly solidified magmas. If granite magma is forced to flow after the biotite has begun to crystallize from it, the flakes of biotite take on a uniform orientation. A batholith that became emplaced during crustal folding generally has a border of primary gneiss whose foliation faithfully follows all the sinuosities of the contact. A primary gneiss formed in this way resembles a gneiss of metamorphic origin, but can be distinguished from it by field and microscopic work.

Quartzite. Quartzite is composed of quartz grains so firmly cemented that when the rock is broken it fractures through the grains, instead of around them. Originally the quartzite was a sandstone, but it has attained its present state (1) as the result of the filling of the pore space of the sandstone by quartz deposited from circulating ground water, or (2) as the result of metamorphism. Quartzite formed by the deposition of a quartz cement is not regarded as a metamorphic rock, however.

Quartzite of metamorphic origin is commonly interbedded with gneiss and mica schist. The original pore space in the sandstone was eliminated mainly by compaction and rearrangement of the quartz already present in the rock. Although quartzite thus formed is intimately interbedded with foliates and has therefore been subjected to the same metamorphic processes that developed the foliation in the inclosing rocks, it rarely shows foliation.

Quartzite is generally a compact hard rock of light color—white, gray, reddish, or buff—and is commonly of vitreous appearance. Such vitreous quartzite is the most durable and resistant of all rocks; and wherever it occurs in notable thickness it has strongly influenced the development of the topography, forming hills or mountains because of its extraordinary resistance to weathering.

Marble. Marble is the metamorphic equivalent of the sedimentary carbonate rocks, limestone and dolomite. During metamorphism the minute grains of which the limestone or dolomite is composed grow in size, with the result that the grains become large enough to be seen by the unaided eye. The resulting marble is harder and more compact through reduction of pore space, and it has purer colors. Some marbles take a good polish. Such visibly crystalline carbonate rocks are, in geologic usage, termed marbles. In commercial practice, however, any carbonate rock that will take a polish is called a marble and is given a trade name.

Most marble is massive, showing no foliation even where it has been subjected to great pressure. As rocks go, marble is highly plastic and therefore flows under moderate differential pressure. If, for example, a dike consisting of rock less plastic than marble is inclosed in marble that was forced to flow under differential pressure, the dike rock, being brittle, was ruptured and torn apart, and the marble flowed in between the dissevered fragments. Marble that has flowed thus in the solid state is an impressive feature in many regions of metamorphic rocks (Fig. 284).

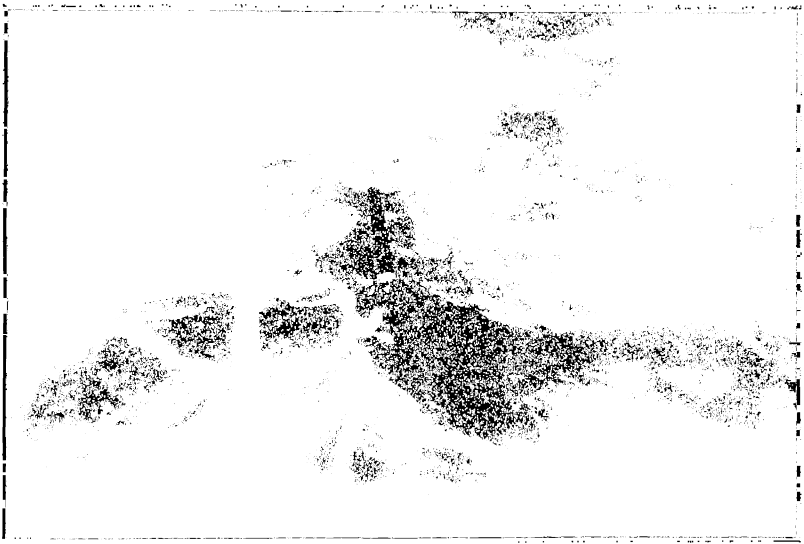


FIG. 284. Irregular dike-like intrusion of white marble into sedimentary beds. Fragments of the dissevered beds are floating in the marble. Near Gouverneur, New York.

Pure marble is white. The mottling, banding, and colors of ornamental varieties are due to impurities: red and yellow tones to oxides of iron, grays and blacks to minor amounts of organic matter. Besides being produced by kinetic metamorphism, marble is also formed by contact metamorphism.

INJECTION METAMORPHISM

Extraordinary effects have been produced in the rocks along the borders of granitic batholiths intrusive into foliates, especially in the zones adjacent to batholiths that were emplaced under a heavy pressure which continued to be active during the cooling and solidification of the batholith. Under such conditions immense numbers of white granite sills are injected into the adjacent rocks parallel to the foliation of these rocks. The granite sills range in thickness from a fraction of an inch upward, and in places are so closely spaced that the resulting composite rock consists half of injected granitic material. The injected material, by its high temperature and its magmatic gases, produces powerful metamorphic effects on the layers of host rock between the sills. The resulting rock has a layered appearance and is called an *injection gneiss.* In places, injection gneisses are remarkably folded and tortuously convoluted, many of the folds being doubled back on themselves. Evidently the whole rock mass was in a highly mobile state when this folding took place, and the injection gneisses thus formed present to us the most astonishing, well-nigh unbelievable sight afforded by any of the rocks of the Earth's crust.

Injection gneisses were formed on an impressive scale during the emplacement of the Coast Range batholith of British Columbia and Alaska. The powerful metamorphic effects of this mighty batholith extend in places as much as 35 miles from its border. The gradual waning in metamorphic power outward from the batholith is shown in diagrammatic perfection in many of the fiords that comprise the waterways of that region. Adjacent to the batholith are injection gneisses, which, as the distance from its border increases, grade into mica schist, then into phyllite, and finally into slate—the whole ensemble an imposing illustration of progressive metamorphism.

POLYMETAMORPHISM

Rocks once metamorphosed may later be subjected to a second metamorphism that differs from the first. In some regions, notably the Alps, the rocks, in fact, have undergone three or more metamorphisms According to conditions, the new changes impressed on the rocks may either advance the metamorphic rank or they may reduce the rank. Such superposed, repeated metamorphisms produce *polymetamorphic rocks.* If the effects of the superposed metamorphism have not been too severe, index minerals of the earlier metamorphism—relicts—can be found even in rock that has been twice or three times metamorphosed, and the successive chapters in its metamorphic history can be deciphered.

A notable form of polymetamorphism is *retrogressive metamorphism,* which is important in interpreting the structural history of many regions.

Retrogressive Metamorphism. Metamorphic rocks, such as the series beginning with slate and comprising phyllite, mica schist, and garnet gneiss, represent stages in progressive metamorphism. All these rocks are of identical chemical composition, but they differ greatly in mineral composition and physical appearance. Each succeeding rock in the series is the product of higher-rank metamorphism than the one that precedes it, the rise in rank being chiefly the result of adjustment to conditions of progressively higher temperature.

A rock adjusted to the condition of highest metamorphic intensity may at some later time, however, be shifted into a new geologic environment, in which the conditions of stability are those of low-rank metamorphic intensity. A garnet gneiss, for example, may be reduced to a phyllite along the base of a great overthrust block of the Earth's crust. Outwardly the resulting phyllite resembles a phyllite formed by progressive metamorphism, but internally, as revealed by the microscope, it still retains evidence that it was formerly a high-rank rock. In short, the high-rank garnet gneiss has retrograded to the low-rank phyllite. By such *retrogressive metamorphism* many varieties of metamorphic rocks have been formed from other metamorphic rocks. By this process the already astonishing diversity of metamorphic rocks has been greatly increased.

The concept of retrogressive metamorphism was first formulated to account for certain metamorphic phenomena seen in the Alps. In general, high-rank metamorphic rocks are formed deep within the crust, whereas low-rank rocks are formed in the upper levels. In places in the Alps, however, high-rank rocks were found to *overlie* low-rank rocks and apparently to grade downward into the underlying low-rank rocks. This anomaly was cleared up by finding that the high-rank rocks had been shoved over the underlying rocks along a great thrust fault. The high-rank rocks in the zone immediately above the fault surface had been transformed into low-rank rocks, so that outwardly they resemble the low-rank rocks below the fault surface. Thus the high-rank rocks seemed to grade downward into low-rank rocks, contrary to the normal relationship. Mere unloading of a region by erosion does not cause the high-rank rocks to be reduced to low-rank rocks. So far as now known, the powerful trigger action of distributed movement induced in rocks being moved along a thrust zone deep in the crust is necessary to bring about retrogressive metamorphism.

Many schists and phyllites in the Alps, the Appalachians, and elsewhere formerly thought to be of normal origin, that is, the results of progressive metamorphism, are in reality the products of retrogressive metamorphism. This new concept, namely that high-rank metamorphic rocks can be reduced to a low-rank state if subjected to the conditions that develop low-rank rocks, is, as we can now see, a logical extension of the fundamental principle of metamorphism that rocks tend to adjust themselves to their environment.

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Note for page 426: In the cores of some mountain chains exposed by erosion, for example the northern portion of the Inyo Range, California, large volumes of rock resembling granite have been formed by replacement of sedimentary rocks. This process whereby rocks have been converted in the solid state by replacement into granitic-looking rocks is called *granitization*. Viewed in the large, the rocks of the Inyo Range look like a bedded formation, but on nearer inspection it is seen that they resemble granite, and that the bedded structure is an inheritance from their sedimentary origin. Granitized rocks constitute a wide aureole around an intrusive granite batholith, which is therefore regarded as having been the source of emanations, gaseous or liquid, which brought in large quantities of material to transform the solid rocks into "granite" and concurrently removed material not required.

In mountain chains more deeply eroded than the Inyo Range, granitization is supposed by many present investigators to have transformed the rocks completely, regardless of their original composition, into granite masses. Thus these investigators account for the great granite batholiths so generally present in the cores of fold mountains (p. 471) The emanations that brought about these radical changes are supposed to have come "from below." These remarkable ideas need more confirmatory evidence before they can be wholly accepted.

In some regions injection gneiss (p. 484), resulting from reaction of the injected magma on the inclosing layers (or septa), has become a new rock type called migmatite (from the Greek word for mixture). If the reaction has been intense, the distinction between the material that was injected and the intervening septa of older rock becomes blurred out, and the resulting rock resembles a homogeneous granite; migmatization is said to grade into granitization.

CHAPTER 18

THE EARTH'S INTERIOR

The materials and conditions that exist at great depths below the Earth's surface can never be known by direct observation. Openings in the crust, such as mines and deep wells, are extremely superficial in comparison with the 4000-mile radius of the globe. We infer that some of the rocks now exposed in the cores of old mountains were at depths of several miles before erosion laid them bare; but even so they were always a part of the "outer shell." If it were possible to make and maintain an opening to the center of the Earth, or to look down with a sort of super X-ray, what would be revealed? Does the composition of material change radically with depth, so that a large part of the interior bears little or no resemblance to the superficial rocks? What is the temperature at various levels? Is any large part of the interior in a fluid condition? How do the materials at great depths behave under the enormous overburden? These are profound questions. They are not prompted by idle curiosity merely; they represent the goal of scientific investigations whose attainment might make clear many of the phenomena seen at the Earth's surface.

Whenever direct evidence is not available, science turns to the indirect and circumstantial. It is possible, with aid of this sort, to draw certain inferences that are sound; but if we seek to advance farther into the unknown we must be content with hypothesis and speculation (Chap. 2). Some suggestions can be accepted as probabilities, with the understanding that they may be found wanting after further investigation. It is well, at the outset, to state clearly the actual basis of fact and to outline the methods for attacking the problem. In this way we shall avoid misconceptions and be able to evaluate each suggestion on its merits.

BASIC DATA AND METHODS

Size and Shape of the Earth. The science of *geodesy*, which is concerned with exact measurement and mapping of the Earth's surface, has determined with precision the dimensions and the form of the globe. This information, obtained by great labor through co-operation of scientists in many countries, is of fundamental importance in problems relating to the Earth as a whole. It is known that the equatorial radius is 3963.4 miles (6378 kilometers), whereas the radius measured along the polar axis is only 3950 miles (6357 kilometers). Since the difference, 13.4 miles, is about $\frac{1}{297}$ of the equatorial radius, it is stated that the *ellipticity* of the Earth is $\frac{1}{297}$. This departure of the globe from the form of a true sphere is a response to rotation. Important inferences are drawn from this known distortion of the Earth, considered with relation to other facts.

Gravity and Density. By precise physical experiments we determine the *constant of gravitation.* The result gives a basis for calculating the total weight of the Earth; and, since its size also is known, it is a simple matter to compute the average density. This value, arrived at by many experimenters, is 5.52; that is, an average sample of the Earth weighs about five and a half times as much as an equal volume of water.

Direct determinations of density of all rocks known to occur at the Earth's surface give an average value of 2.7. As this is less than half the density of the whole Earth, the interior must consist of much heavier material than the outer part. Other inferences are discussed in a later paragraph.

Relation between Density and Form. In detail, the surface of the Earth is highly irregular. Continental masses stand well above deep-sea floors; plateaus, mountains, and deep troughs make irregularities of smaller order (p. 5). On a small globe made to true scale these surface features appear insignificant; but in actual dimensions some of them are large, and from a human viewpoint the irregularity is of the utmost importance since without it the sea would be worldwide.

Theoretically the surface of the Earth would be level if the material below the surface were uniform in character. In reality we know that the rocks exposed to observation differ considerably in composition and in density. Basalt and other dark-colored igneous rocks are notably heavier than granite. By geologic investigation and by highly technical instrumental determinations a strong probability has been established that the rocks underlying the deep sea are in general denser than those composing the continental masses (Chap. 2). Moreover it is concluded from careful geodetic study that the great mountain ranges of the Earth are composed of and underlain by material that is slightly deficient in density compared with the crust as a whole. These facts suggest that the larger surface irregularities are not haphazard, but have a fundamental cause; that large differences in surface elevation are directly related to differences in the density of the underlying rocks. Important inferences based on this relationship are discussed in Chapter 2.

Behavior toward the Moon and the Sun. The Moon and the Sun exert a constant pull on the Earth, and yielding of the sea water to this pull gives rise to the tides. By careful and ingenious experiments it has been determined that the body of the Earth also responds to the tidal force, but the yielding is very minute—about what would be expected if the Earth were composed of the strongest steel.

Another effect is produced by solar and lunar attraction on the equatorial bulge of the Earth, causing the globe to wabble slowly as it spins. Consequently the north pole of the heavens shifts slowly from year to year. This effect, called *precession,* is known precisely, and the forces involved can be calculated closely. The use of this information is discussed below.

Response of the Earth to Seismic Waves. The most reliable information about the Earth's interior is furnished by study of earthquake waves (p. 407). Vibrations that are transmitted through the body of the Earth and recorded by seismographs at varying distances from an earthquake focus reveal the elastic properties at different depths. This is a most promising field of study, and undoubtedly many important revelations will be made through continued researches in seismology. At present the most significant facts resulting from the study are the following:

(1) The *long* waves, which move around the Earth in the rocks directly beneath the surface (p. 407), travel at a higher speed through the deep-sea floors than through the continental masses. For example, a seismographic record received in Japan of an earthquake in California shows a higher speed for the long waves than a record of the same shock received in New York. We conclude that granite, the predominant kind of rock in the continental masses, does not form the deep-sea floors.[1] This conclusion is also reached independently from geologic evidence, which suggests strongly that basalt and similar dark, heavy rocks predominate under the deep seas (p. 14).

(2) Both the primary and the secondary waves traveling through the Earth (p. 407) increase in speed with increasing depth of the path, down to a depth of about 1800 miles (2900 kilometers). Thus, in Fig. 285, the primary and secondary waves that reach station *c* travel more

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[1] The same kind of evidence suggests that the floors of the Atlantic and Pacific oceans differ somewhat from each other in composition.

rapidly than those received at *b*, which in turn move at a higher average speed than those that emerge at *a*. The rate of increase is shown in the following partial table.

Distance of Receiving Station from Earthquake Focus, in Degrees	Velocity, in Miles per Second	
	Primary Wave	Secondary Wave
30° (*a*, Fig. 285)	5.4	3.0
60° (*b*, Fig. 285)	6.8	3.7
90° (*c*, Fig. 285)	7.9	4.3

An important cause of the higher wave velocities at greater depth is compression by gravity, which increases the rigidity and consequently the elasticity of the material. It is highly significant that the secondary, or transverse, waves are transmitted freely. Waves of this nature move only through rigid substances. Therefore modern seismology has disproved the old concept that below a comparatively thin crust the body of the Earth consists of hot liquid.

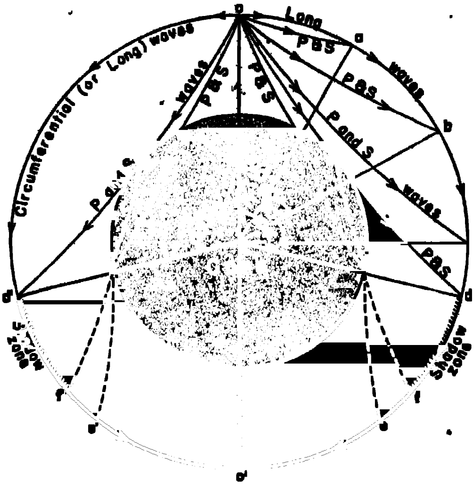

FIG. 285. Section through the Earth, to illustrate behavior of earthquake waves, originating beneath *o*. Stations between *o* and *d*, and between *o* and *d'*, receive a complete record. At stations beyond *d* and *d'* (as *e*, *f*, *e'*, *f'*), record of the secondary wave is missing or at best very faint. The primary wave is refracted in its passage to these stations, as shown (*one*, *olf*, *ome'*, *okf'*); as a result, a zone beyond *d* and another beyond *d'* (*df* and *d'f'*) receive no record of any through-waves. Long waves, moving around the circumference in both directions from *o*, reach all stations.

The statement that the velocities of primary and secondary waves increase continuously downward requires one important modification, on the basis of a recent discovery by seismologists. At a depth of about

50 miles (80 kilometers) below the Earth's surface the velocities of these waves fall off noticeably, and the slower rates of travel hold to a depth of nearly 100 miles (150 kilometers), where the increase in velocities is resumed. Thus a zone about 45 miles (70 kilometers) thick, with its upper limit 50 miles beneath us, has distinctive properties. Possibly the rock in this zone is near the melting temperature, under the pressures at that depth, and as a result the elastic properties of the rock are affected. If this is the correct explanation, the zone may be of great importance in the mechanism for maintaining isostatic equilibrium of the crust (p. 18) and in connection with igneous activity.

(3) Below a depth of about 1800 miles (2900 kilometers) the behavior of the elastic waves changes abruptly; the speed of the primary wave drops from 8 miles to about 5 miles per second, and the secondary wave either disappears or becomes very faint. Furthermore the primary wave penetrating to greater depth is refracted or bent, just as a ray of light is bent in passing from air into water (Fig. 285). As evidence of these changes in the seismic waves, stations located farther than 104 degrees from the epicenter of an earthquake get at best only a very weak record of secondary vibrations; and in a belt of considerable width directly beyond the 104-degree limit the primary wave is lost, owing to refraction (Fig. 285); this belt is known to seismologists as the "shadow zone." It is evident, therefore, that the Earth has a core, with a radius of more than 2100 miles (3400 kilometers) which differs radically in chemical composition or in physical condition, or in both respects, from the thick shell that surrounds it.

(4) There is strong evidence also that the thick shell itself is divided into two major parts. Although the velocity of the waves continues to increase down to 1800 miles, at a depth of about 600 miles (1000 kilometers) the *rate* of increase abruptly falls off. It is probable, therefore, that the kind of material below the 600-mile level is different from that above.

(5) The central core and the change at a depth of 600 miles are detected by records of earthquakes at a great distance. "Near earthquakes"—that is, shocks that occur within a few hundred miles of a recording station—furnish important information about the shallow zones of the earth. In addition to the ordinary primary and secondary tremors, the record of a near earthquake usually shows an additional set of preliminary waves that have been refracted from a deeper zone of higher density (Fig. 286). In this evidence there is a definite indication that the rocks with which we are familiar at the surface form a comparatively thin layer, resting on rock of a distinctly heavier type. Ac-

cording to some calculations this layer is less than 10 miles thick; but values that have been obtained range up to 35 or 40 miles. These figures are not necessarily contradictory, since the layer probably varies considerably in thickness from place to place.[1]

Temperatures in the Earth. Volcanoes and hot springs indicate high temperatures at depth locally, and direct measurements in deep mines and wells suggest a universal increase in temperature downward. The rate of this increase varies between wide limits. In some places it is as high as 1°F. in 30 feet; in other places, as in the deep gold mines of the Transvaal, South Africa, it is only 1°F. in 250 feet. The average for all observations is about 1°F. in 60 feet. At a depth of slightly more than 16,000 feet in a well drilled for oil in California, the measured temperature was 400°F.

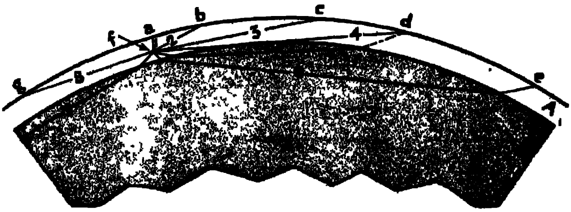

FIG. 286. Explanation of the evidence indicating an outer shell (*A*) resting on a zone of denser rock (*B*). Waves radiating out from a shallow earthquake focus (*f*) follow the paths, 1, 2, 3, 4, 5, 6, etc., and reach recording stations. Station *d* receives the ordinary *P* and *S* waves over route 4, and also a second pair of preliminary tremors over route 5; this pair has been refracted down into a denser zone and refracted up again as shown.

Records from several stations near *d*, some closer to *f*, and others farther away, give a basis for estimating the thickness of the shell *A*.

Vertical scale greatly exaggerated.

The determination in recent years that all known rocks contain appreciable amounts of the radioactive elements uranium and thorium is of great importance (p. 26). It appears certain that these materials are not present in considerable quantities below a comparatively shallow outer zone of the Earth. Radioactive substances break down at a constant rate, and the disintegration liberates heat. Since rocks conduct heat away at an extremely slow rate, it can be calculated that the presence of uranium or thorium in minute traces, even at moderate depths, would result in permanent fusion. Granites contain higher percentages of radioactive substances than other known rocks, and hence it is argued that granites, which are the predominant rocks in the continental masses, are limited to shallow depth.

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[1] Some records of near earthquakes suggest a second thin layer, of intermediate density. Possibly some of the calculations for thickness treat the two layers as one; but further evidence is required to establish the existence of the intermediate layer.

Behavior of Rocks under Pressure. Since ordinary rocks are brittle, they fracture and crush under high pressures in the laboratory unless special precautions are taken. When a rock specimen is confined on all sides and subjected to strong compression, equal in all directions, its size decreases slightly; in other words rock is somewhat compressible. On release of the pressure the specimen tends to recover its original size, by elastic expansion.

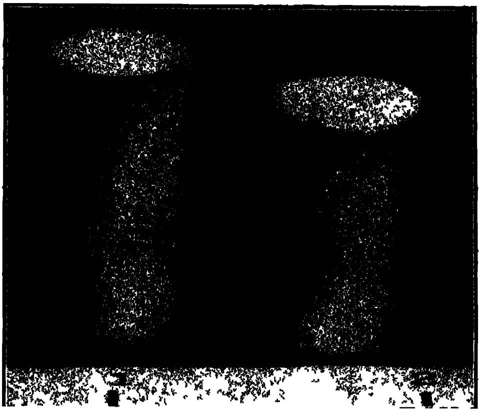

FIG. 287. Result of a laboratory test by D. T. Griggs. The cylinder, made of fine-grained limestone, initially had the dimensions shown at left. It was placed in a specially constructed press, and pressure up to 185,000 pounds per square inch was applied at the ends of the cylinder, while liquid around the sides of the test piece maintained a uniform confining pressure of 150,000 pounds per square inch. The cylinder was shortened and thickened by plastic flow, to the dimensions shown at right. Scale has 1-inch divisions.

A limestone cylinder placed in a specially constructed hydraulic press, and subjected to pressure that is great on all sides but strongest at the ends, changes its form by growing shorter and thicker. Up to a certain value of pressure, known as the elastic limit of the rock for the given conditions, the specimen will regain its original dimensions when the pressure is removed. But if, when the elastic limit is reached, the pressure on the ends of the cylinder is increased and the experiment continued for some time, the shortening and thickening become permanent (Fig. 287). Testing and microscopic examination of the deformed cylinder reveals no loss of strength and no fracturing; under the unbalanced pressure the rock has *flowed* slowly. Weak materials such as beeswax can be molded easily with the hand; these are known as *plastic* substances. Strong rock also is plastic under sufficiently intense confining pressure.

Since rocks deep in the Earth are confined under high pressure, there is little doubt that under certain conditions these rocks are deformed by plastic flow. The depth necessary for this kind of deformation should be greater for strong rocks like granite than for weak rocks such a mudstone. For any kind of rock the ease of plastic flow increases wit rising temperature; thus the effect of heat is to lower the strength rocks (p. 414).

INFERENCE AND HYPOTHESIS

Constitution of the Earth. The outer part of the Earth consists of material much lighter than the average of the Earth as a whole (p. 438). Based on this fact alone, however, several assumptions might be made as to the arrangement of light and heavy substances between the surface and the center. For example, one theory might be that light rocks, like granite, form a shell a few miles or tens of miles thick, and that below this shell the Earth is composed of heavy substance with a uniform density of approximately 6. Again it might be assumed that from the outer zone of low density there is a gradual and progressive increase to a density of 9 or 10 at the center. By either of these arrangements the average density of 5.52 could result. Besides these two suggestions a number of other assumptions as to distribution of density might be made, all of them consistent with the known average density.

Fortunately there are other checks to guide us in attacking the problem. Any assumed distribution of the density must harmonize with the mathematical and mechanical knowledge to which reference has been made above (pp. 437–444) and also with the evidence from seismology. The small degree of flattening at the poles of the Earth suggests convincingly that a large percentage of the mass is concentrated in the central portion; it is calculated that if the outer part had nearly the average density, 5.52, the centrifugal force would produce an ellipticity much greater than $\frac{1}{297}$. Moreover, if the density were uniformly distributed the equatorial bulge would contain so much mass that the precessional effect, from solar and lunar attraction, would be much larger than it is. These considerations therefore favor the inference that density increases slowly downward in the Earth, and that the central core is composed of exceptionally heavy material.

Once this point is reached in the inquiry, another problem presents itself. Is the increase in density toward the center of the Earth due wholly to compression under enormous weight, or is there a concentration, toward the center, of metals that normally are heavy? Some students of the problem have argued that compression of ordinary rock material is a sufficient explanation. At a depth of 1 mile each square foot of rock bears a weight of 450 tons; and with each additional mile the pressure is increased by more than this amount, as the material grows progressively denser. Near the center, pressures amount to more than 2 million tons per square foot. Without question such intense pressure has an effect in compressing the material. Since pressures that are possible in laboratory experiments do not approach pressures that obtain near the center of the Earth, we can not state positively how much compression can result. However, from various lines of evidence and reasoning it appears unlikely that ordinary rocks can be compressed sufficiently to give the high average density that the Earth possesses. Especially significant is the abrupt drop, at a depth of 600 miles, in the rate at which the speed of earthquake waves increases downward (p. 441). If the entire Earth were made of ordinary rock, the speed of transmission of such waves should increase with depth at a nearly uniform rate, as long as the material remains rigid. Seismic waves testify that the Earth is rigid to a depth of at least 1800 miles, and therefore the material between the 600- and 1800-mile levels can not be ordinary rock. Hence it is logical to infer also that the more pronounced break indicated by seismic evidence at a depth of 1800 miles represents an abrupt change in composition. Therefore most scientists favor the view that the core of the Earth is metallic.

Which of the metals is most likely to exist in such abundance in the Earth? The most satisfactory answer to this question comes from consideration of the material that reaches us from outer space. The majority of known meteorites are composed of iron and nickel, and the others consist of dark-colored, heavy rock. Accordingly it is commonly inferred that the central core, more than 2100 miles in radius and therefore occupying more than ⅛ the volume of the Earth, is composed chiefly of nickel-iron with an average density of about 8, probably made even higher under the compression to which the core is subject (Fig. 288).

Emphasis has been placed on meteorites as samples of cosmic material and probably representative of our planet. The thoughtful student will sense an apparent discrepancy in the fact that the majority of known meteorites are metallic, whereas metals do not appear to dominate the composition of the Earth. Astronomers have learned, however, from spectroscopic study of meteors as they flame in the sky, that a preponderance of this foreign material coming into our atmosphere consists of common elements contained in our dark igneous rocks. Actually, therefore, stony material in these bodies is more abundant than metals. Metallic meteorites found at the Earth's surface are in the majority because they are more distinctive than stony meteorites, which resemble some of our mundane rocks. Moreover the minerals in stony meteorites decompose rather easily, whereas the nickel-iron bodies are more resistant to weathering. Thus there is no real discrepancy between the supposed composition of the Earth and the average composition of meteorites.

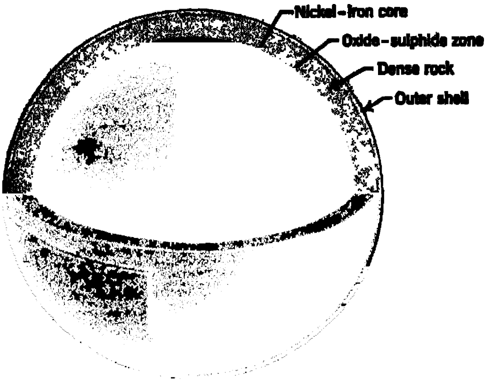

FIG. 288. The principal zones of the Earth and their inferred composition. Thickness of the outer shell is shown to exaggerated scale.

The sum of evidence indicates convincingly that the Earth has a "density stratification," with the heaviest material near the center and the lightest near the surface. This arrangement is one of the strongest reasons for believing that the Earth passed through a molten stage in the early part of its history. During this stage, gravity caused most of the heavy metal to settle toward the center, and progressively lighter substances formed spherical shells between the core and the surface. A useful analogy is furnished by a smelter furnace in which metals are extracted from ore. When a large mass of ore is smelted the metal settles to the bottom of the container; directly above is a layer of sulphides and oxides which, although they are heavy, are distinctly lighter than the metal; and at the top is slag made of the molten rock. By reasoning from analogy, it is inferred that the shell, 1200 miles thick, extending downward from a depth of 600 miles to the metallic core, is composed of sulphides and oxides with an average density of 5[illegible] (Fig. 288). This inference finds support in the presence of sulphides and oxides as common constituents of meteorites, and it explains satisfactorily the decline in the rate of increase of seismic-wave velocities below the 600-mile level.

The seismologic evidence suggests that rock extends continuously from the surface to a depth of 600 miles. Probably the greater part of this thick shell consists of dark-colored rock of a type rarely seen at the surface; its density is 4 or slightly less. Above this zone is a layer, a few tens of miles thick, composed of the familiar rock types; gabbro at the bottom of the shell grades upward through diorite into the granite that forms the continental masses.

Figure 288 shows diagrammatically this inferred constitution of the Earth. Probably the shells grade into each other and so are not delimited as sharply as the drawing suggests. It should be kept in mind that the entire concept is a hypothesis, no part of which is subject to direct proof at present; it is merely an attempt to give a picture that is consistent with all known facts. However, the array of facts is becoming formidable, and therefore the hypothesis is infinitely more valuable than the uncontrolled guesses of the last century.

Temperatures at Depth. The average change in temperature in the Earth for a given unit distance is known as the *geothermal gradient.* If the gradient determined in mines and boreholes should continue downward unchanged, the temperature at the center would exceed 350,000°F.; but for several reasons the average rate of change in the shallow zone can not be used with confidence for great depths The length of the deepest opening used in estimating the gradient is about $1/1300$ the Earth's radius; and the value obtained for this thin skin can not be accepted with any confidence for the whole body. Rocks such as granite are extremely poor conductors of heat. Therefore if the temperature at a depth of several miles should be high, say 1000°C., heat would flow out very slowly, and the change in temperature for each 100 feet would be considerable. It is altogether likely that the heat conductivity improves with depth, both because compression of the rocks increases and because metallic substances, which are notably good conductors, probably grow more important in the deeper zones. Therefore the gradient in the shallow zone probably is much larger than for any other part of the globe and can not be used to calculate the temperatures at great depth.

It is probable also that much of the heat conducted to the Earth's surface and lost by radiation does not come from great depth, but is generated in a shallow zone. The distribution in all known rocks of uranium and other radioactive elements has been mentioned (p. 26).

These elements disintegrate slowly but continuously, with evolution of heat. The possible significance of this process in connection with igneous activity has been discussed (Chap. 14).

Whatever may be the true value for the average temperature gradient, many lines of evidence suggest that temperatures in the interior are high. Probably they are above the melting points for the materials under surface conditions; but, since the earthquake waves testify that no considerable part of the globe is molten in the outer 1800 miles of its radius, it is certain that heat is not great enough in this portion to cause general liquefaction under the great pressures that prevail. However, at depths of a few tens of miles the temperatures may be so high that the rocks have very little strength. Even though high pressures maintain the rigidity that is indicated by fast-moving earthquake waves, rock that is "potentially liquid" from high temperature would flow under long-continued stresses. This concept is of great importance in considering the adjustments of the crust under heavy loads (pp. 17–19).

There are still differences of opinion concerning the physical condition of the central core. Thus far no undoubted record of secondary waves has been detected at stations situated more than 104 degrees from an earthquake focus (that is, at stations beyond *d* and *d'*, Fig. 285); this evidence from seismograms suggests a fluid central core, since the transverse elastic waves are transmitted only through rigid material. Many seismologists accept the evidence as conclusive, and speak confidently of a fluid core, more than 4000 miles in diameter. However, other students have suggested that the secondary waves, which are of longer period than the primary, are almost totally reflected within the central core and so are almost or entirely missing in records at the antipodes. In any case, a positive determination that the central core is or is not rigid must await further study.

Relation to Surface Features. The nature and behavior of materials inside the Earth concern us chiefly in connection with our search for the explanation of major phenomena at the Earth's surface. In particular, the great folds and faults that characterize mountain belts appear to reflect activities in the invisible interior of the Earth. Repeatedly in geologic history the Earth's crust has been buckled and riven by mountain-making forces, and the deformed rocks have been forced up in the building of mighty chains, such as the Alps, the Himalayas, and the Andes. What is the origin of these forces, and how have they acted in fashioning the great relief features of the globe?

Any attempt to answer these difficult questions must start with consideration of the evidence furnished by the mountain belts themselves. This evidence is discussed in Chapter 19.

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CHAPTER 19

THE ORIGIN AND HISTORY OF MOUNTAINS

Mountains are of great importance in geology, since they furnish a large part of the information on which the science is based. Dynamic processes, such as stream erosion and glaciation, are especially vigorous, and their effects are strikingly evident, in high ranges. The elevation of rock masses to great heights has resulted in dissection to unusual depths; consequently a mountain region affords excellent opportunity for descriptive study of rock formations and for deciphering the history they record. But although mountains give aid in solving many problems relating to the Earth, they also present mysteries in themselves. Why have sea floors of remote periods become the lofty highlands of today? What generates the enormous forces that bend, break, and mash the rocks in mountain zones? These and other questions on mountain genesis still await satisfactory answers; but the architectural features of great ranges at least offer hints as to their origin and are in themselves fascinating subjects of study.

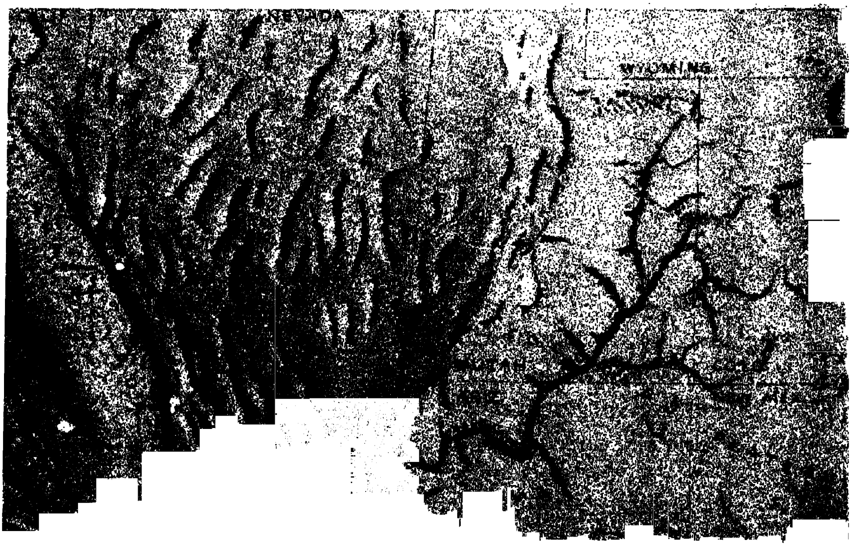
MOUNTAIN UNITS

An isolated high land mass that rises above comparatively low surroundings is described simply as a *mountain.* Examples are Stone Mountain in Georgia, Mount Monadnock in New Hampshire, and Mount Etna in Sicily. No arbitrary lower limit is set to the height of features that are called mountains; in a low plains country the term is applied by the inhabitants to steep hills only 300 or 400 feet high, whereas in the Rocky Mountain region and in other rugged districts some relief features 2000 feet or more in height are known locally as hills or buttes. Thus there is some vagueness in the definition; but an essential characteristic of mountains, aside from great height, is comparatively limited width at the top. Near the mouth of the Grand Canyon, in northwestern Arizona, an observer looking eastward faces a steep, rugged escarpment 4000 feet high (Fig. 289). This precipitous rise in the land surface has a mountainous appearance, and it is sometimes referred to locally as "the mountain." However, if one climbs to the top of the steep escarpment he looks eastward for tens of miles across nearly level ground. Therefore the high escarpment is the edge of an extensive *plateau* and not a mountain front.

Generally mountain masses do not stand alone but are parts of distinct units that vary in size and plan, from irregular *groups,* like the La Sal Mountains of Utah and the Adirondack Mountains of northern New York, to the enormous belt that extends more or less regularly from the Pyrenees eastward across Europe and Asia to the East Indies Descriptions of the larger units or their parts employ somewhat loosely the terms *range, system,* and *chain.* As it is desirable to use descriptive terms with a definite meaning, the usage proposed many years ago by J. D. Dana is followed here.

A *mountain range* is either a single large, complex ridge or a series of clearly related ridges that make a fairly continuous and compact unit Excellent types are the Sierra Nevada in eastern California (Fig. 289) and the Front Range of Colorado. A group of ranges that are similar in their general form, structure, and alignment, and presumably owe their origin to the same general causes, constitutes a *mountain system.*

Fig. 289. Mountains and plateaus of southwestern United States. The principal units at this latitude are, from east to west, the Southern Rocky Mountains, the Colorado Plateau, the Basin-and-Range province, the Sierra Nevada, and the Coast Ranges. Numerous details omitted to avoid obscuring of major units.



Thus the Basin Range system (Fig. 289) in Nevada and adjoining States consists of many distinct ranges that trend north or northwest and have similar structure and form. The Rocky Mountain system is a great assemblage of ranges, made at approximately the same time, extending from near the Mexican boundary northward through the United States and western Canada. The term *mountain chain* is used somewhat more loosely, to designate any elongate unit consisting of several ranges or groups, regardless of similarity in form or age relationships.

But a still more comprehensive term is needed to refer to a series of chains or systems that make a more or less unified belt of vast extent. For this purpose a Spanish word was borrowed by the famous traveler Humboldt. All the mountain units in western North America, from the eastern border of the Rocky Mountains to the Pacific coast, are known collectively as the North American *Cordillera*[1] (Fig. 289). Similarly the entire broad mountain belt that extends almost continuously from Alaska to Cape Horn is known as the American *cordilleras*. However, the same term has not been adopted universally for the major mountain belts of the Earth. In the literature the great mountain unit of southern Europe and Asia is designated variously as the *Mediterranean* (or *Eurasiatic*) *chains, zone,* or *belt* of mountains.

ORIGIN OF MOUNTAINS

Mountains owe their origin to various causes, and differences in the structure and the plan of mountain units are due largely to this fact. The principal agents involved are volcanic activity, movements of the crust, and differential erosion. Commonly two or more processes combine to produce complex results. However, the discussion will be clarified by a classification that recognizes the dominant processes.

VOLCANIC MOUNTAINS

The most obvious process of mountain making is the accumulation of lavas and other volcanic products into heaps of mountainous size. Some of the loftiest peaks in the world, such as Chimborazo (20,517 feet) and Aconcagua (23,393 feet) in the Andes, and Kilimanjaro (19,321 feet) in Africa, have been built directly by volcanic action. Many such peaks, however, have their bases on high plateaus, and therefore the actual height due to volcanism is much less than the altitudes above sealevel suggest. A large number of oceanic islands, like those of the Hawaiian group, are great volcanic piles, and some of them appear to be seated directly on the deep-sea floors (p. 304).

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[1] Pronounced either kôr-dĭl-yā'rȧ or kŏr-dĭl'ẽr-ȧ.

Some volcanic mountains, such as Vesuvius, are isolated cones rising above comparatively flat country. Other high cones occur in closely associated groups, like the San Francisco Mountains on the plateau south of the Grand Canyon in Arizona; elsewhere they are arranged in fairly regular rows, along great faults or other lines of weakness in the crust. Mountain masses formed directly by igneous extrusion are sometimes called "mountains of accumulation." In areas of widespread volcanism the volcanic materials build also extensive plateaus, such as the Absaroka Plateau east of Yellowstone Park and the great Columbia Plateau of Washington and Oregon. After a time stream erosion carves such plateaus into high ridges and peaks, which thus become mountain groups that are the joint product of accumulation and erosion. Erosion by streams or glaciers has modified the forms of all volcanic mountains, particularly those in which the volcanic activity has been for a long time extinct.

PLATEAU REMNANTS

High plateaus are dissected by stream erosion, and during a late stage in the process some of the more favored residuals have sufficient height, in relation to their surroundings, to be called mountains. Many of the larger buttes in western United States are examples. The Catskill Mountains in New York are remnants of a former extensive high plateau, the greater part of which has been removed by fluvial erosion. A clear indication of this old plateau is given by the nearly horizontal sedimentary formations of which the Catskills are made; the thick strata now end abruptly at the steep borders of the mountains, and obviously their original extent must have been much greater. An early stage in a similar development is to be seen in the Grand Canyon of the Colorado, where great pyramidal erosion remnants rise from the depths of the chasm (Fig. 290). These masses are dwarfed by the high plateau surrounding them; but if they could be placed on a plain they would dominate the landscape. It is a logical expectation that the present youthful plateau eventually will be dissected thoroughly by the Colorado and its tributaries, and will then be represented only by scattered high residuals separated by plains and valleys.

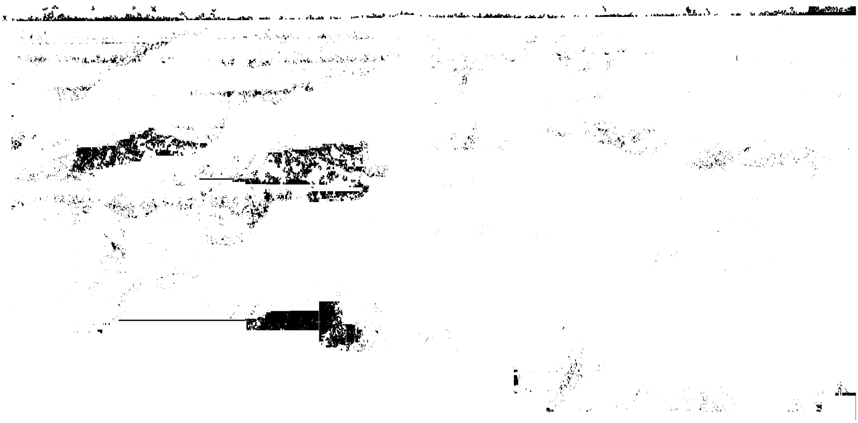
High plateau remnants are also called "mountains of erosion." However, large-scale differential erosion is made possible only by great uplift, and therefore residuals of mountainous size are the joint product of crustal movement and erosion. The forms and the distribution of the residuals are determined largely by the kinds and the structure of the rocks in the plateau. Horizontal sedimentary formations or homogeneous igneous rocks are dissected by dendritic stream patterns, and remnants owe their preservation to chance location on divides or near the headwaters of major streams. In some plateaus, however, the rocks are steeply tilted strata whose edges were cut down to a nearly uniform level at a low elevation by previous erosion (Fig. 299). When such a peneplaned surface is warped up to form a plateau, the weak rocks, especially the shales and limestones, are attacked more effectively by weathering and streams than are the edges of resistant sandstone strata, which therefore are left standing in relief (Fig. 300). This aspect of mountain history is discussed at more length on pages 500–501.

MOUNTAINS WHOSE STRUCTURE REFLECTS CRUSTAL MOVEMENTS

Volcanic mountains and plateau remnants are important geologically, since they form groups of considerable size in the world today and probably have had wide distribution during past geologic periods. However, in most of the dominating mountain units now in existence

DRAWN BY W. H. HOLMES.

Fig 290. Part of the Grand Canyon seen from Point Sublime, showing dissection of the plateau into isolated remnants of mountainous size. The far rim of the Canyon is about 10 miles from the foreground, and tops of the two great pyramids are more than 4000 feet above the river.



the major relief has been determined either directly or indirectly by localized crustal movements that have caused more or less severe disturbance of the rocks. In some of the young mountain systems much of the local relief was caused directly by these movements; in belts of mountains which have experienced deep erosion and perhaps more than one regional upwarping, the original disturbance of the rocks is important chiefly in guiding erosion. But even if much of the present mountain relief is due chiefly to differential erosion, any characteristic structure produced by crustal movement has large importance in classifying mountain units. According to the nature of the movements, as indicated in the resulting structure and form, mountains of this general type are divided into four classes. (1) Faulting on a large scale results in relative uplift of crustal blocks, with or without tilting. Ranges whose structure is produced chiefly by this process are *fault-block mountains*. (2) Some vertical movements result in arching of the rocks into a general domal form, either nearly circular or somewhat elliptical in plan. *Dome mountains* result from this process. (3) More commonly the forces deforming the crust produce large anticlines and synclines, giving rise to the structure of *fold mountains*. (4) In most of the great mountain belts we see the combined effects of two or more types of movement, particularly folding and faulting, with complications produced by igneous intrusion. The resulting mountains are of the *complex* type, but local sections may be classified according to the process that has played a dominant role.

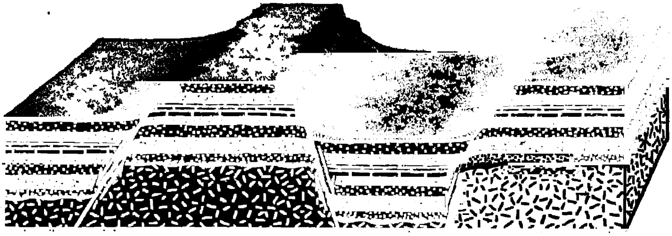
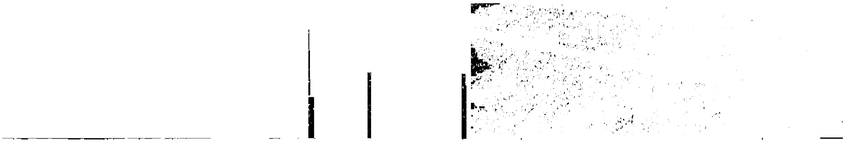


FIG. 291. Mountain ranges formed by simple normal faulting. The horsts form the ranges, the grabens the intermont basins.

Excellent examples of each mountain type exist; but Nature is complex, and consequently various combinations of the different types are most common. The influence of erosion in varying degree is evident in all mountains, regardless of type.



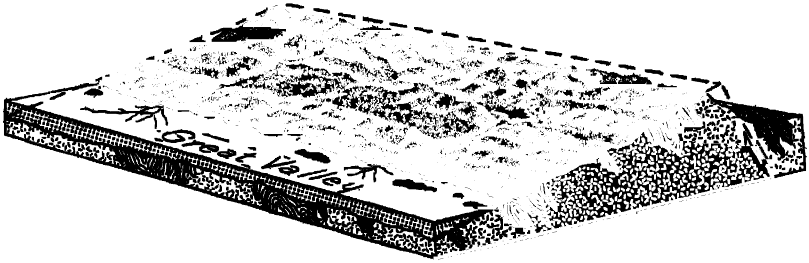
SPENCE AIR PHOTOS.

Fig. 292. Front of a fault-block mountain range, south of Goodsprings, Nevada. The Union Pacific Railroad tracks (shown in foreground) are near the base of a slope built up of debris eroded from the range. A large fault is concealed near the top of this slope. The highest peak is about 3000 feet above the foreground.

Fault-Block Mountains

Assume that a system of intersecting fractures, reaching to great depth, divides part of the Earth's crust into blocks or masses of very large dimensions. Then conceivably mountains can be formed directly by movements of these blocks in several ways. (*a*) If the region is initially a high plateau, some of the blocks may be depressed several thousand feet, leaving other blocks in their original elevated positions, to form mountain ranges. (*b*) Regardless of original altitudes, some of the blocks may be elevated to mountain heights, by forces acting largely in the vertical direction, leaving adjacent blocks relatively depressed. (*c*) All the blocks may move downward or upward, but differentially so that in the end some stand much higher than others. (*d*) Each of several blocks may be tilted or rotated, one edge being elevated and the opposite edge depressed. Whatever the nature of the movement, ranges that are formed by faulting are termed *fault-block mountains* (Figs. 291, 292, 293). Movements on both normal and reverse faults give rise to such mountains.

Fault blocks as we actually see them have been more or less modified by erosion. Debris worn from the high masses tends to bury those at low elevations (Figs. 291, 292). In time this combination of erosion and deposition nearly or quite obliterates the mountain relief, especially in a region of interior drainage where nearly all the debris is caught in basins between the mountains (p. 114). At a later date, as a result of broad regional uplift accompanied by change of climate, or because of important changes in the drainage of adjoining regions, the waste may be removed from the original intermont troughs and the ranges given a new lease of life. Obviously such resurrected ranges are the direct result of differential erosion, and strictly they are plateau remnants. However, the original faulting was the chief factor in guiding erosion and continues to be reflected in the mountain forms.



Modified from F. E. Matthes.

FIG. 293. Part of the Sierra Nevada, California, looking northeast, showing the general form and structure. The east front is an eroded fault scarp; the gentle western slope follows the tilt of the uplifted block. Remnants of folded rocks engulfed in granite record ancient crustal movements and enormous igneous intrusions, after which long-continued erosion formed a peneplane. In late geologic time the present range was formed by tilting uplift along the great fault. Glaciation shaped the rugged peaks at the summit and enlarged the valleys on the western slope.

Vertical scale considerably exaggerated. Broken lines suggest approximate form of block before dissection by erosion and burial under sediments in Great Valley.

The Sierra Nevada of California is a tilted crust-block 400 miles long, 40 to 60 miles wide (Figs. 289, 293). Its eastern edge has been uplifted 2 miles or more, to form an abrupt scarp facing eastward. Roads from the east ascend this precipitous front with difficulty; but west of the crest they descend on a long, gentle slope—the tilted upper surface of the block. In the Great Valley of California, sediments thousands of feet deep have accumulated on the depressed portion of the rotated mass. A great series of fault mountains in Nevada and parts of neighboring States forms the Basin Range system (p. 452; Fig. 289). The great dislocations responsible for these ranges and for the Sierra Nevada do not represent the first disturbance of the region. In earlier periods the thick sedimentary rocks in Nevada and eastern California were folded and thrust-faulted, and great igneous masses were intruded into them. Mountains that existed soon after those ancient events have long since disappeared, and in late geologic time the old deformed crust was broken by great faults, to form the present generation of mountains. Probably some of the ranges are still growing, for, within historic time, movements have occurred on several of the faults, giving rise to violent earthquakes. As a large part of the region has no drainage to the sea, the mountains are partly buried by accumulations of their own debris.

Fault-block mountains in various stages of destruction are found in northern Africa, in Arabia, in central Asia, in Japan, and in many other parts of the world. The thick red sandstones of Connecticut and Massachusetts are broken by great faults, and the resulting blocks have a strong tilt eastward; but the mountains that probably were formed by these dislocations disappeared through erosion long ago. Similar structural relics of ancient fault-block mountains, representing various geologic periods, are widely distributed in all continents

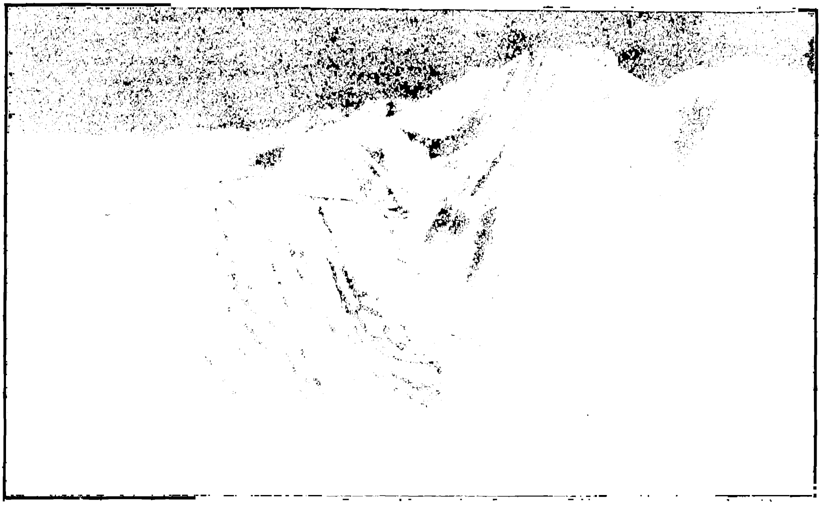
Dome Mountains

Mountains whose structure reflects crustal uplift of distinctly domal form may be classed together, regardless of size or the exact cause of the uplift. The simplest are laccolithic domes, made by the bowing up of strata above thick, lens-shaped intrusions of molten rock (p. 287). Ordinarily a dome of this kind that is high enough to be called a mountain has lost more or less of its original cover through erosion; and not uncommonly the resistant igneous mass, almost completely denuded, stands within circular or elliptical ridges formed by the upturned edges of the more resistant strata (Fig. 196, p. 288). There are several excellent examples in the vicinity of the Black Hills. The Henry Mountains of Utah, classic examples of the type, are a large group of laccoliths and related igneous bodies in various stages of dissection by erosion. But not all mountains of this kind have ideally simple structure. In the Moccasin Mountains of Montana the intruding magma ruptured the covering strata and lifted them irregularly.

Some cylindrical masses of igneous material (stocks, p. 292), rising vertically from great depths, have failed to reach the surface but at their upper limits have bent sedimentary beds strongly upward to form symmetrical domes. In early stages of dissection, mountains of this kind are indistinguishable from laccolithic domes. When the cylindrical mass has been eroded to a lower level, however, it is seen that the mass cuts across the strata with which it is in contact. Some of the larger igneous bodies in the Henry Mountains and in the Black Hills appear to have this form, although superficially they resemble laccoliths.

Fold and Complex Mountains

Mountains in which rocks are strongly folded and broken are commonly described according to their internal structure, whether or not the deforming forces were the direct cause of present high altitudes.



Geological Survey of Canada.

Fig. 294. Mount Perdrix, Alberta. After the strata were folded, erosion removed vast quantities of rock, and a closely compressed syncline, originally a low part of the folded section, now appears in one of the high mountain ridges.

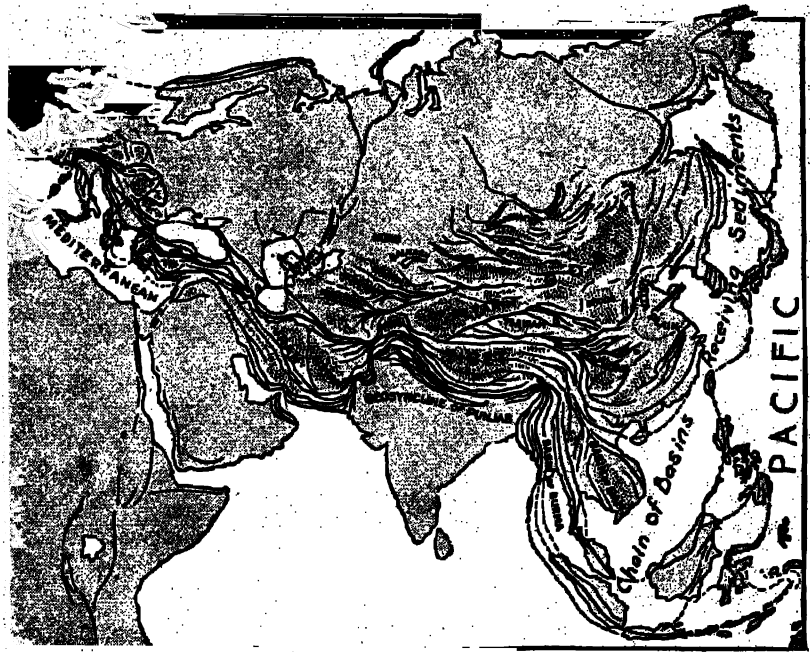
Some old mountain units are strictly remnants of erosion; however, it is evident from their structure that a certain type of crustal deformation attended their early development, and their structural characteristics are used as the basis of their classification. For example, the Appalachians of eastern North America and the Cape ranges of South Africa have had long and varied careers. The original highlands were eroded through long ages, and they almost or quite disappeared; and the present ridges have been brought into relief by differential erosion as the result of later upwarping of the areas containing the old mountain "roots." In these and other mountain units, numerous synclines, which originally were the lowest parts of the folded belt in which they occur, now are found in high peaks or ridges (Fig. 294). It is well to keep these examples in mind throughout the discussion that follows.

Deformed rocks are characteristic of all great mountain zones, but it is not to be taken for granted that the deformation gave rise directly to the present mountain heights. Mountain *structure* and mountain *elevation* may not have any direct relation to each other. Nevertheless the structure continues as a dominant factor in determining relief because it guides differential erosion.

All the great mountain chains of the Earth include folded sedimentary rocks as a conspicuous part of their structure. These chains, therefore, are sometimes classed together as *fold* mountains, although faulting, igneous intrusion, and other important processes besides simple folding have played some part in their origin. Actually every great system is more or less *complex* in its structure; but certain mountain units exhibit fairly regular folding of rock formations as an outstanding structural characteristic. The Jura Mountains in Switzerland and parts of the Appalachian Mountains are excellent examples. The Rocky Mountains and the Alps, both characterized by enormous thrust faults in addition to folds, are outstanding illustrations of complex units. However, many parts of the Appalachians also are complicated by thrust faulting; and, as there are all stages of gradation between the simpler sections of this chain and the almost incredible complexity of the Alps, it is clear that *fold* and *complex* mountains can not be separated as sharply contrasted structural types.

General Character of Fold and Complex Mountains. From examination of a globe or a world map it is apparent that each of the prominent mountain belts is composed of numerous ranges disposed somewhat irregularly, but in long stretches with the same general orientation. Some of the ranges are nearly straight in plan; but many are strongly curved into the form of great bows or arcs. The Alps, Carpathians, Himalayas, and other chains of Eurasia are striking examples of this *arcuate* type (Fig. 295).

An important part of the bedrock exposed in each of these mountain belts consists of distorted sedimentary formations. Most of these strata, now on the flanks or even on the highest summits of the ranges, represent deposits in former seas, on deltas, and in swamps bordering the sea. Owing to the strong folding and faulting of these strata, followed by planation and dissection through erosion, the full thickness of the sedimentary cover can be seen and measured. In some mountain belts these thicknesses are astonishing; 4, 5, or even 6 miles are by no means exceptional values, and in some mountain areas the total sedimentary sections exceed 40,000 feet. It will occur to some readers that similar thicknesses may be common also outside of mountain zones, but are not known because not exposed by erosion. However, natural exposures, supplemented by well records, indicate clearly that sedimentary formations grow conspicuously thinner away from a folded mountain belt. Thus the strata in the Appalachians average 20,000 feet or more in thickness along the central axis of the folded tract, with a maximum thickness near 40,000 feet; but at no great distance to the west the thickness is less than 10,000 feet, and in the Mississippi Valley it is only 4000 to 5000 feet (Fig. 297). On the east side of the Appalachians this series of sedimentary strata is not recognized, and it will be shown presently that the deposits never extended indefinitely eastward from their present limit. Therefore the excessively thick sedimentary sections occupy a long and relatively narrow belt that corresponds closely in trend to the axis of the chain. This general relationship exists also in the Rockies, the Mediterranean chains, and other



Modified from Berkey and Morris, American Museum of Natural History.

FIG. 295. Map of Asia, with parts of Europe and northeastern Africa. Trends of the principal mountain chains of Asia and southern Europe shown by black lines. The arcuate form of many units is striking. The trends are generally parallel with the south and east coasts of Asia. Note the series of great marine basins receiving modern sediments between the east coast and the island chains.

great mountain belts. It is a logical conclusion that accumulation of stratified rocks in abnormal thicknesses had a significant connection with the development of each of these mountain units. Therefore these thick accumulations merit particular attention.

Development of Geosynclines. Sedimentary strata in the great mountain belts consist of conglomerates and sandstones interbedded with shales and limestones. Because they are thick deposits and contain coarse sediments, they must have been laid down near the margins of lands that underwent prolonged erosion. The great thickness of deposits may suggest that sedimentation began in an excessively deep basin. However, there is unquestionable evidence that most of the sediments involved in mountain folds were laid down in shallow water or at only moderate depths. Accumulation of such deposits to a total thickness of several miles indicates that slow subsidence of the sea floor was continuous or recurrent while deposition was in progress. Moreover, as enormous volumes of sediments were delivered into the subsiding basin during long periods, a wasting land mass must have risen continuously or recurrently adjacent to the area of sedimentation. An

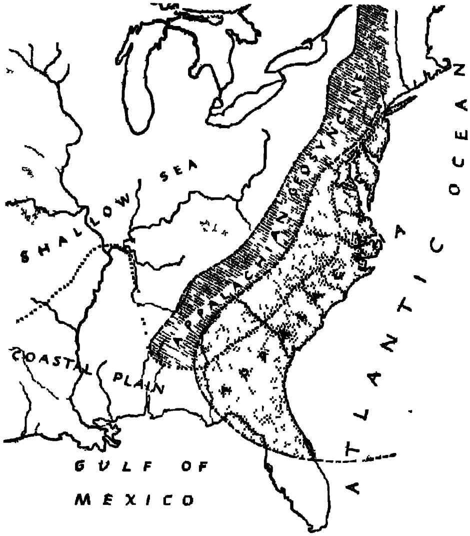


FIG. 296. Map showing general location of the Appalachian geosyncline and of the old land which furnished a large part of the sediments that were deposited in the sinking trough. Relations in New England uncertain. Dotted line shows inner margin of younger deposits that mantle the present Coastal Plain.

elongate subsiding tract of this nature, which receives thick deposits of sediments during subsidence, is known as a geosyncline (p. 370; Figs. 296, 297). A modern example on a large scale may be the great chain of shallow seas parallel to the east coast of Asia, into which the Hwang Ho and other great rivers of the continent as well as shorter streams from the western slopes of the Japanese and other islands, are pouring

Fig. 297

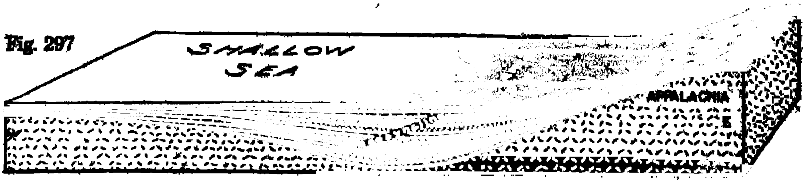


Fig. 298

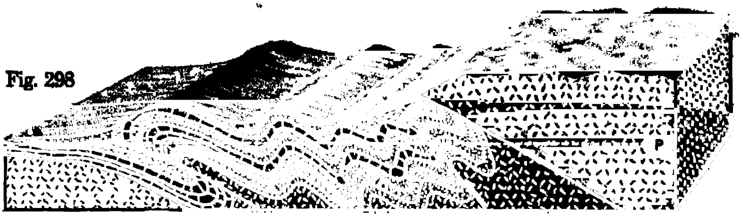


Fig. 299

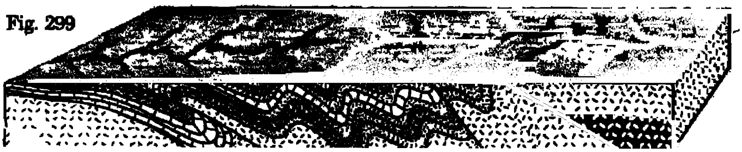
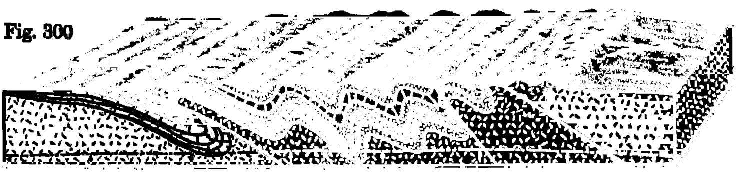


Fig. 300



FIGS. 297–300. Four stages in the evolution of the present Appalachians. Views looking north. Each block about 200 miles long. Vertical scale exaggerated.

FIG. 297. The geosyncline receiving sediments from a former land on the east.

FIG. 298. Folding and thrust faulting of the rocks in the geosyncline. The form and height of the surface are entirely hypothetical; it is not known how much uplift resulted directly from the folding or to what extent erosion kept pace with the uplift. After several geologic periods had elapsed the region was worn to a peneplane, indicated by the broken line at P.

FIG. 299. General appearance of the peneplaned Appalachian region near the close of the Mesozoic era. The eastern part of the region was submerged and received a veneer of coastal-plain sediments; possibly this veneer covered much of the folded belt, and has been removed by erosion.

FIG. 300. During Cenozoic time the region was warped up, as indicated by arrows at the bottom of the block; belts of weak sedimentary rock were eroded faster than the resistant strata, which therefore make the present ridges. (The broken line corresponds to the [illegible] of the block in Fig. 299.)

floods of sediment (Fig. 295). Another example, on a smaller scale, is a belt of thick sediments including the Mississippi delta and extending eastward along the north side of the Gulf of Mexico to northern Florida.

In the Appalachians various features of the strata indicate that conditions within the old geosynclinal trough fluctuated repeatedly. Sandstones and shales with abundant ripple marks and mud cracks are interbedded with thick limestones that contain marine fossils. Such relations imply a shifting shoreline and considerable variation in depth of water. In fact at some periods the sea gave place to great delta plains or to enormous swamps in which materials for coal beds accumulated. These changes depended on the relative rates of subsidence and sedimentation. If sinking of the trough halted for a considerable time, accumulating sediments made the sea shallow or even displaced the sea water entirely over wide areas. With renewal of subsidence the water came back. If the adjacent land was elevated rapidly for a time, erosion may have been stimulated sufficiently to keep the seaway full even though subsidence of the trough was continuous.

From study of some sedimentary sections in the Appalachians it is clear that the coarser sediments are on the east, and that they grade westward into marine shales and limestones (Fig. 297). Therefore the land from which these sediments were eroded lay east of the geosyncline.[1] A narrow belt of ancient rocks near the present coast may represent the western edge of the former land; but how far that land extended eastward, over the area of the present continental shelf, is not known. Although the vast quantities of sediments it supplied may suggest considerable width, even a narrow land that was rising almost continuously may have been adequate. The name *Appalachia* is applied to this ancient land of unknown extent (Fig. 296). The sea that lay west of it, covering much of the present Mississippi Valley region, was shallow and had shorelines that shifted widely during its long history.

Possibly Appalachia was not a unit land mass, as suggested in Fig. 296, but consisted of great island chains near the border of the continent, similar to chains now flanking the east coast of Asia. The Asiatic chains are partly submerged mountain ranges, some of which developed from folding of thick sedimentary deposits. We may speculate, therefore, that in remote geologic ages a geosyncline developed near the present eastern coastline of North America; that deformation of this ancient geosyncline formed mountains which grew into island chains; and that these rising islands were the source of sediments laid down in a younger basin on the west, the Appalachian geosyncline. The great volumes of sediments testify clearly to the former existence of Appalachia, but do not reveal its exact form or its full extent.

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[1] Land areas on the west also contributed some deposits to the geosyncline, at certain stages of its development. The exact location and extent of these land areas, and the amounts of sediment they supplied, have not been determined.

Other great mountain systems, in both of the hemispheres, have had beginnings similar to that of the Appalachians. The Rocky Mountains are on the site of a great geosyncline that stretched from the Gulf of Mexico to the Arctic, with highlands west of it; strata folded in the Caucasus Mountains are made of sediments that were derived from lands to the south while seas stretched northward over Russia. The time occupied in the accumulation of sediments in a geosyncline has ranged from tens of millions to hundreds of millions of years. The Appalachian geosyncline was a basin of deposition during six geologic periods, nearly 300 million years in total duration.

Stages of Severe Crustal Movement. The structural features found in each of the great mountain belts indicate that crumpling of the old geosyncline and its burden of sediments was performed by forces acting in a lateral direction, parallel to the Earth's surface. Thus in zones of most intensive folding, not only are the folds closed so that their limbs are parallel, but also they are even more severely compressed, with mashing of the beds and the production of very complicated structures. This is illustrated by sections in the Appalachians (Fig. 300) and in the Alps.

Some strong folding movements occurred while development of the Appalachian geosyncline was in progress. In eastern Pennsylvania, in the Hudson Valley, and in southeastern Canada the strata deposited during the early stages of geosynclinal history were thrown into steep folds and now lie with strong angular unconformity beneath formations that were laid down later. Further strong disturbance, however, occurred after the youngest of the geosynclinal sediments were deposited (Fig. 298). Likewise on the floor of the sea in which the Alpine formations accumulated there were repeated disturbances that folded and broke the strata, and we see the result in numerous unconformities. As the folding continued the entire belt finally emerged from the sea, and the final crushing and thrusting by lateral compression occurred after nearly all the sedimentary strata had been formed.

Thrust Faulting. It is to be expected that such extreme folding of rocks would result in rupturing and displacement of the strata along great faults. We find, accordingly, that faulting is especially common

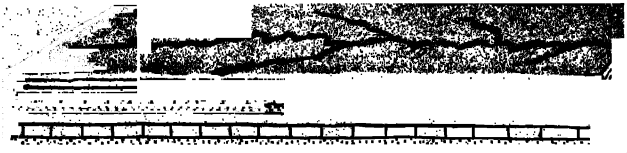

Fig. 301

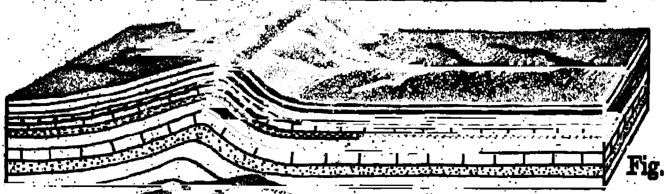

Fig. 302

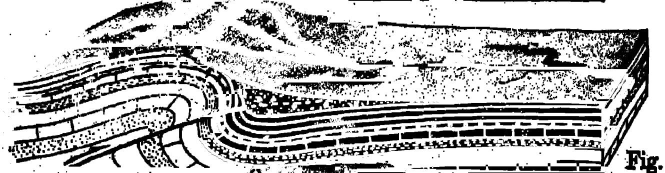

Fig. 303

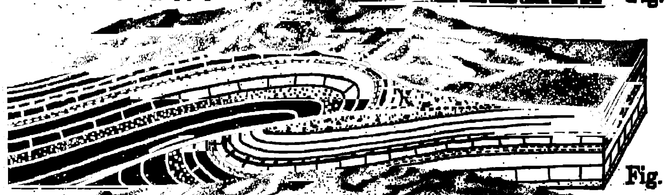

Fig. 304

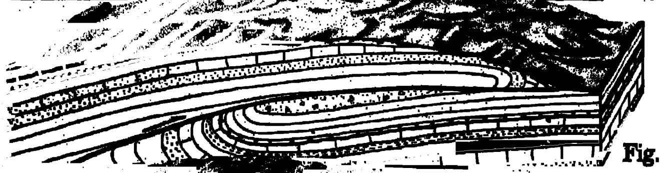

Fig. 305

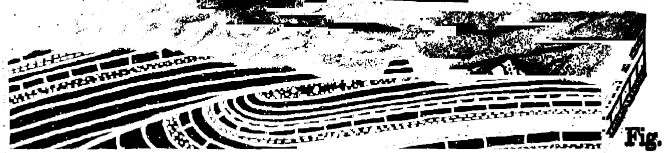

Fig. 306

FIGS. 301–306. Development of a great thrust fault. Length of block about 50 miles.

FIG. 301. Undisturbed strata before the movement has begun.

FIG. 302. Horizontal compression has formed a large anticline.

FIG. 303. The fold has developed into a thrust fault. Debris eroded from the rising structure has accumulated in the synclinal depression at the base of the range.

FIG. 304. The mass is pushed out over the land surface and overrides some of its own alluvial debris.

FIG. 305. The thrust reaches its maximum extent.

FIG. 306. After movement has ceased erosion has full sway. The thrust mass is cut down and back. Note the isolated remnants of the older rocks. (Compare Fig. 307.)

in mountain ranges. As we pass from consideration of the simpler fold ranges to those of more complex types, the faulting becomes more pronounced until finally it reaches extreme development in thrust faults of great magnitude (Figs. 301–306). An excellent example is the great Lewis thrust in Glacier National Park, Montana (Fig. 307). The low dips of the thrusts and their trends parallel with the axes of the ranges indicate that the thrusts were made by horizontal compression, just as were the folds. In fact many of the folds, after being overturned, were broken, and the breaks developed into thrusts; thus the origin of the two types of structure from the same forces is clearly established.

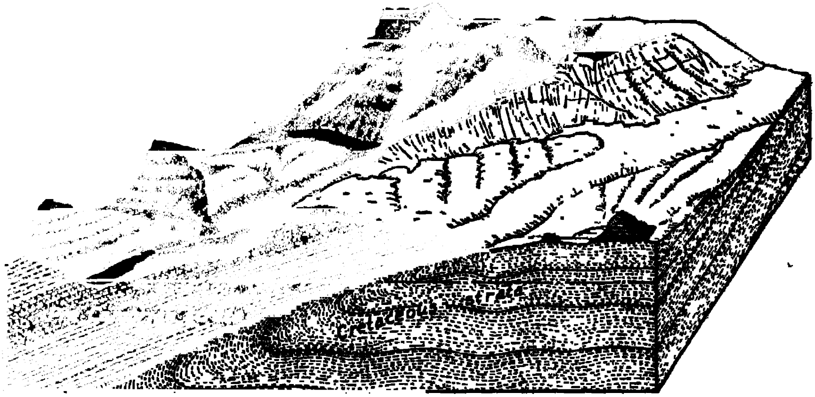

FIG. 307. Part of the great Lewis thrust, in Glacier National Park, Montana. The old rocks pushed up and forward in the thrust are indicated, both at the surface and beneath, by the darker color. The thrust has been cut back farthest where the largest stream has eroded its valley. Chief Mountain (*C*) and another small remnant are "mountains without roots"; they show that the thrust reached farther east than at present, but probably do not record its maximum extent. (Compare Fig. 306.)

View looking northwest. Front of block about 10 miles long.

Amount of Compression. The magnitude of the forces and of the masses involved is indicated by the amount of shortening that has been produced in some of the great folded belts. It is estimated that in the Appalachians the original width of the geosyncline was decreased at least 40 or 50 miles, and in some sections more, by the crumpling of the mass and the thrust faulting. Thus if the folded strata in Pennsylvania, which now resemble a crumpled blanket, could be smoothed out toward the southeast, their extent would be increased sufficiently to cover most of the State of New Jersey. In the Rocky Mountains also the structure represents shortening by tens of miles, and the amount of compression in the Alps is much greater.

Variations of Fold Structure. It is to be expected that the results of folding should differ considerably in character between the several mountain systems, or between distant parts of the same system. Differences in thickness of sediments, in proportions between strong and weak formations, and in severity of the lateral forces are reflected in the individuality of mountain folds. The Jura Mountains, a small member of the Alpine system, have irregular folds, many of them broken and crumpled. These folds were produced far out in front of the Alps proper, in a relatively thin sheet of strata, as an incidental effect of the forces that deformed the greater Alpine zone (pp. 470, 472). The Appalachians present a wider variety of fold structures. In the slate and anthracite regions of eastern Pennsylvania the folds are closely compressed, many of them to the isoclinal form (p. 368; Fig. 236), and the axial planes are strongly overturned toward the northwest. Farther west in the State the folds tend to be open and upright; and the deformation dies out westward. In the mountain belt farther south, throughout Virginia, Tennessee, and northern Alabama, many of the folds were ruptured by the severe compression and developed into thrust faults (Figs. 301–306). This kind of complexity is especially pronounced in the Alps, which merit special mention.

Thrust Faults and Recumbent Folds of the Alps. Alpine structure is characterized by great folds that have been pushed over to a horizontal attitude, and by flat thrusts that are related to these overturned folds. These features are developed on an unparalleled scale, with the result that the Alps consist of a series of great rock sheets, driven one over another and overlapping like the shingles on a roof. The Germans call the individual sheets *Decken;* the French refer to them as *nappes.*

Because of their location, the Alps have received more intensive study than any other mountains. Accordingly, in spite of astonishing complexity, their structure and history are fairly well known. As in the Appalachians, the deformed rocks in the Alps include thick marine deposits; but a larger part of the Alpine sediments bears evidence of deposition in deep water, far from any shore. Crustal movement began in Mesozoic time, with pressure from the direction of Africa. The layers of sediments on the sea floor were bowed up slowly, until chains of islands appeared above sealevel. During early Cenozoic time the compression accelerated powerfully, and an enormous rock sheet was driven northward over the geosyncline. Beneath this sheet the weak sedimentary formations were intensely crumpled. With recurrent thrusting other sheets were driven forward, and all were severely folded.

Erosion cut valleys through the sheets (Fig. 308), exposing the entire series; and in parts of the Alpine area nearly the whole of one or more sheets has been swept away, leaving remnants of old rocks to form isolated peaks standing on younger rocks that were overridden and covered during the thrusting movement. Like Chief Mountain, Montana (Fig. 307), isolated peaks that have this anomalous relation are "mountains without roots." The Mythen, prominent peaks south of Lake Zurich, are famous examples. Some of these masses are 50 miles or more north of their original positions. Heim, the great Swiss master of Alpine structure, tells us that the Alpine zone as a whole was made narrower by nearly 200 miles owing to the thrusting and folding. Locally, as in the Simplon Tunnel section, the original width was reduced as much as 90 per cent.

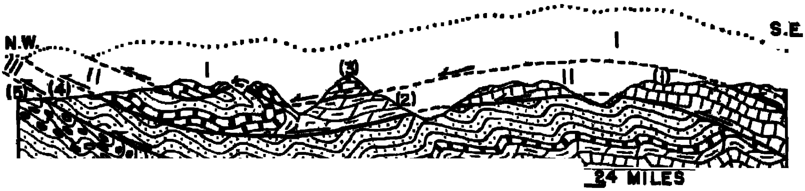

FIG. 308. Generalized section across part of the Swiss Alps. Three great thrust sheets, or *nappes*, are numbered I, II, III. Sedimentary formations are numbered 1 to 5, in order from oldest to youngest. Dotted line suggests original form of thrust sheet I, which has been largely removed by erosion. Isolated mass under 3 is a "mountain without roots."

Depth of Folding. How deep is the segment of the crust affected by folding, thrusting, and mashing in mountain zones? There is of course much uncertainty about this matter, but for some localities a partial answer can be given. In the Alps, erosion has cut through the folded and uplifted cover of strata in several places, exposing the older rocks on which the sediments were deposited. Mont Blanc and the Aar *massif* are parts of the floor of the former geosyncline, now lifted to a high level and completely denuded. These areas were not in the deepest parts of the old basin; yet they were covered with sediments many thousands of feet thick, and the crustal movements reached much deeper than the present surface, since the old rocks are sliced with thrusts and otherwise deformed. The Appalachians have not been elevated sufficiently to permit erosion to expose the floor in the deepest part of the old trough; but thrust faults have brought some of the lowest sedimentary formations to the level of the present surface. Similar evidence from the Rocky Mountains and from the older mountain zones of Scotland and Scandinavia leaves no doubt that deformation by horizontal compression extends through a depth of at least several miles, but how much deeper the effects may reach is not known. On the other hand it is certain that some folding ends at comparatively shallow depth. The folds of the Jura Mountains, northwest of the Alps, are closely compressed; but the sedimentary formations involved in the folding have a total thickness less than a mile, and the older rocks beneath were not affected by the deformation. Near the bottom of the sedimentary section there is a weak formation of shale. When the Alpine zone was being crushed by powerful thrusts from the south the deformed rocks, crowded toward the north, transmitted part of the pressure to the flat-lying strata beyond. The weak shale at the base of the section served as a lubricant, permitting the overlying strata to wrinkle into close folds by sliding over the "basement" rocks (Fig. 309). To duplicate the effect in miniature, lay a thin block of tissue paper on a table with the far edge of the block against a book, and push against the near edge; the paper yields by wrinkling into folds, at the same time sliding on the table. Shallow folds of the Jura type appear to have developed incidentally in connection with more important and deeper-seated crustal movements (Fig. 310).

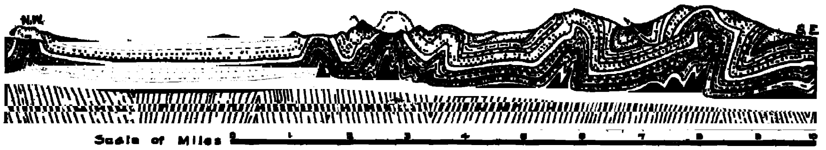

FIG. 309. Section through part of the Jura Mountains. Thrust from the southeast (right) caused the plate of stratified rocks to fold by slipping over the older rocks beneath. (See Fig. 310.)

Mountain Elevation. There is a natural tendency to assume that all the excess material crowded into a mountain zone by folding and thrusting was forced above the general level, and that growth of the great ranges in height ended as soon as horizontal compression ceased. For a long time, indeed, this conclusion was taken for granted, and attempts were made to compute the original heights of eroded ranges by restoring the folds from study of the eroded limbs. As the steps in mountain history become clearer, however, it is found that much of the actual elevation occurred at a time distinctly later than the folding and thrusting. After the Rocky Mountain deformation in Mesozoic and early Cenozoic time, the folded and faulted area was eroded to a nearly even surface at a low altitude; and the present great heights in the Rockies are due to strong upwarping at a later date. Similarly, after much of the thrusting and folding was complete, the Alps had only moderate height, and the sea washed the flanks of the ranges both, on the north and on the south. In late geologic time, powerful upwarping of the entire mountain belt carried the Alpine summits to great heights. The Andes and the Himalayas have had similar histories.

Therefore horizontal compressive forces, important though they have been in the development of mountain structure, are not a sufficient explanation of mountain height. What other forces may be involved? A possible answer to this question is reserved for later paragraphs.

Role of Igneous Agencies. Although some great ranges contain little or no visible igneous rock, nevertheless intrusion and extrusion of igneous material commonly have been associated with mountain genesis. Great batholiths have been intruded into many belts of folding and thrusting. A granitic mass of this kind that has become exposed by deep erosion is spoken of as the "granite core" of a range (Fig. 198, p. 290). Intrusion of hot magma, combined with the folding and mashing of the older rocks, has caused profound metamorphic effects over wide areas (p. 419). One of the finest examples of such batholithic masses is in the Coast Range of western Canada, where granitic rocks are exposed in a continuous belt more than 1200 miles long (p. 291).

Intrusions of molted magma have not only made batholiths at the cores of ranges, but also have pressed upward into the folded and faulted strata to form sills, laccoliths, and dikes (Chap. 13). In many mountain zones the magmas have broken through to the surface on a grand scale, as evidenced by the vast accumulations of volcanic breccia, lavas, and other volcanic materials near Yellowstone National Park, along the Andean chains, and in numerous other mountain belts the world over.

Since igneous activity involves tremendous energy, it may seem a logical suggestion that igneous processes have supplied the force required for folding and thrusting. However, evidence of igneous action is wholly lacking in some long belts of intense deformation. Moreover, many intrusive masses in mountain belts were emplaced after the folding and thrusting had been accomplished, since the igneous bodies cut across the folds and faults. Therefore the igneous activity was more probably an effect than a cause of the deformation.

Possible Interpretation of Mountain History. Why are thick geosynclinal deposits commonly involved in mountain deformation? What causes great uplift in mountain zones long after folding and thrusting movements have ended? Why have great batholiths formed beneath mountain belts? These questions can not be answered with confidence at present, but some suggestions find support in the accumulating evidence. Figure 310 illustrates some of these suggestions. In (*a*), the geosyncline *G* has received vast quantities of sediments from a land mass *L*, which was rising persistently while the geosynclinal tract was sinking. This rising land was at least in part a growing mountain belt, subject to repeated deformation from horizontal pressure in the crust Thus mountain-making forces were already in action while the geosyncline was forming; in fact, every geosyncline seems to have been one manifestation of these forces. Although the accumulating weight of sediments was no doubt a factor in causing subsidence of the geosynclinal floor, the sediments were of much lower density than the rock displaced at depth (Fig. 6, p. 18), and their weight alone could not have kept the tract below sealevel, by causing a mile of sinking for every mile in thickness of added deposits. The mechanism responsible for horizontal compression in the crust may have caused downbending beneath the geosynclinal tract. Under this suggestion, the sediments began accumulating at *G* because there was a ready-made basin to receive them, and they continued to pile up because progressive down-

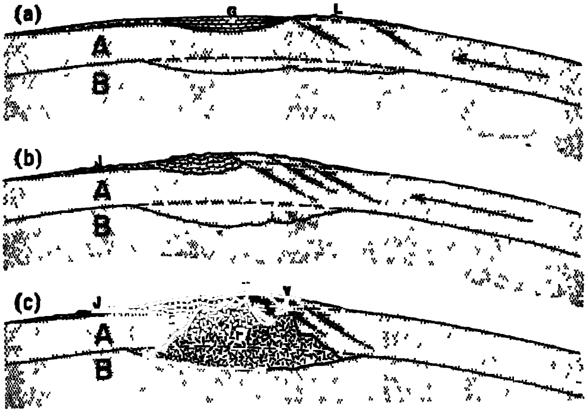

FIG. 310. Hypothetical sections through a mountain belt, showing three stages of development, (*a*), (*b*), (*c*). *A*, outer shell made of low-density (continental) rock; *B*, deeper zone made of denser rock. Broken lines indicate normal thickness of low-density shell, before thickening by deformation. Arrows in (*a*) and (*b*) indicate lateral compression, with no suggestion of its origin. In (*a*), downbending beneath geosyncline *G* presumably is caused by forces that deform the shell beneath the rising land *L*. In (*b*) the deformed and thickened shell is in isostatic balance. In (*c*) fusion has formed batholithic masses *F*, and volcanic material *V* is extruded. *J*, superficial folds like those of the Jura Mountains (Fig. 309) in thin sedimentary section outside the geosyncline. (Vertical scale much exaggerated.)

bending of the crust maintained the catchment basin. Weight of the sediments accentuated the downbending in some degree, but was not a primary cause of subsidence.

Figure 310 (*b*) represents renewed deformation of the mountain belt, including the entire width of the geosynclinal tract. Compression has thickened the outer light shell (*A*) of the crust, and the top of the thickened portion has risen to higher altitude in conformity with the principle of isostatic balance (Chap. 2). At a later stage, (*c*) rising temperature in the lower part of the thickened shell has caused large-scale fusion, with development of a batholith that cuts across folds and faults. Expansion has caused increased uplift, and vigorous erosion has carved the high surface into rugged forms. A logical cause of the large-scale fusion may have been accumulation of heat from disintegration of radioactive elements (p. 335), which are more abundant in the outer granitic shell *A* than in the dark-colored rocks of the deeper zone *B*. Since granitic rocks fuse at lower temperatures than basaltic or gabbroic rocks, the resulting igneous activity would be confined to the thickened part of the granitic shell *A*. Starting near the base of this shell, fusion would proceed upward until temperature in the upper part of the fused mass fell below the temperature required for further melting. Dissipation of heat through volcanic activity may have been one factor in halting the upward progress of batholithic invasion.

The succession of events outlined here is oversimplified, in comparison with the complete records found in many mountain districts. Thus in parts of the Rocky Mountains there were two or more episodes of volcanic activity, each probably corresponding to a spreading of igneous fusion in the underlying crust. Each renewal of major igneous activity in turn may have been connected with a distinct episode of compressive deformation, of which several occurred in the Rocky Mountain region, separated by long intervals of time. Mountain histories have been varied and complex; hence a simple generalization can not be made for all parts of any one mountain system, much less for all great systems. Moreover, the hypothetical nature of interpretations suggested above, and in Fig. 310, must be emphasized. The problems are too vast, and evidence now available is too meager, to permit more than intelligent guesses about the mechanism of mountain building.

Later Stages of Mountain History. As long as the compressive forces are at work, a mountain system grows in so far as its structure is concerned. Whether it actually rises in height or not depends on the adjustment between (1) vertical movements that tend to make it rise, and (2) the work of erosion which tends to cut it down. Always during the formative period this struggle goes on, and the height of a range at any time is the resultant of these two forces. When crustal movements cease erosion has full sway. After an enormous lapse of time the mountain forms may be obliterated; but even then some clue to their former existence will be furnished by the upturned and dislocated nature of the eroded strata (Fig. 311), by the widespread metamorphism of the rocks, by the faults that cut them, and by the presence of large granitic intrusive masses. We can not determine the former altitudes, for, as LeConte has said, "We find only the bones of the extinct mountains"; but these remains indicate the trends and extent of the ancient ranges. Thus from the kinds and attitudes of the rocks of southern New England, which is now only a hilly country, we are led to infer that it was once a mountainous region, with ranges trending generally north and south.

Crustal movements in the mountain zone do not cease, however, at the time the ranges reach their maximum height. As erosion proceeds, its work is partly offset and the mountain forms are complicated by recurrent upwarping of the region. The old Appalachians have been bowed up repeatedly, even in late geologic time. After their birth late in the Paleozoic era they experienced long-continued erosion; and although uplifts probably occurred, the entire folded tract finally was reduced to a peneplane. During more recent epochs the region was

N. H. DARTON, U. S. GEOLOGICAL SURVEY.

Fig. 311. East front of the Laramie Mountains, Wyoming. Erosion has reduced these mountains to low elevation; but even if they should eventually be eroded to a plain, the upturned strata of limestone, sandstone, and shale will mark the position and the trend of the old range.

warped up strongly and dissected, and therefore the present mountain ridges are strictly plateau remnants (Chap. 20). The repeated uplifts are logically explained as movements to restore isostatic balance after removal of great load through prolonged erosion (Fig. 6, p. 18).

DATING OF MOVEMENTS IN A MOUNTAIN ZONE

To fix the geologic period in which the sedimentary formations of a geosyncline were deformed, it is necessary to determine the age not only of the youngest strata involved in the disturbance but also of the oldest strata that were deposited later (Fig. 312). The closeness of dating by this method depends upon the length of the interval between deposition of the two sets of formations. Thus if the youngest deformed rocks (for example, *A*, Fig. 312) are of early Jurassic age, and the oldest undisturbed formation (as *B*, Fig. 312) was deposited early in the Cenozoic, the movement may have occurred either in the Jurassic or in the Cretaceous period (see Appendix D). The latest Appalachian folding is dated as Permian, since late Pennsylvanian rocks are affected and Triassic rocks are not.

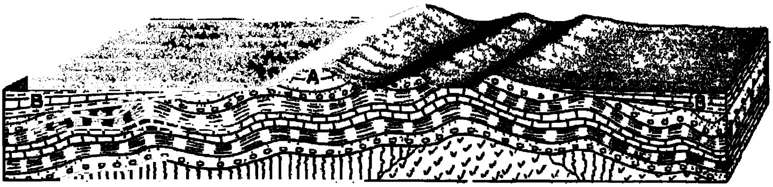

FIG. 312. Cross-section of a range with fold structure, showing evidence of the date at which deformation occurred. *A*, youngest surviving strata that were involved in the folding; the deformation occurred after they were deposited. *B*, strata deposited after the deformation. Length of section about 50 miles.

In some mountains the rocks have been disturbed in several periods, and therefore the method just explained can be used to date only the latest movement. However, the earlier disturbances generally are recorded by unconformities (p. 386). If a surface of unconformity cuts across strongly folded rocks, and the formations above the unconformity also are deformed, there were at least two important disturbances. The several pulses of violent deformation that affected the Alpine zone are differentiated by evidence of this kind.

Mountain-making movements of any kind are dated by following the same general principles. The latest episode of uplift on faults can not be older than the youngest rocks affected by the faulting.

The Ultimate Cause of Crustal Movements

The mechanism of crustal deformation and mountain uplift represented in Fig. 310 is largely hypothetical. However, it does not attempt to explain the origin of forces that cause the deformation. This fundamental mystery about the Earth has engaged the attention of scientists through several generations, but the problem is still unsolved. Some of the current hypotheses hold great interest, but they can be given only brief mention here.

Origin of Steep Faults. Some steep faults that bound mountain blocks may have resulted from strains set up by movements to restore isostatic balance disturbed by erosion and deposition (Chap. 2). Other faults appear to be related to irregular shifting of magmas at depth, during the emplacement of igneous bodies or the extrusion of volcanic materials. Many dome mountains owe their formation to igneous intrusion, and therefore the problem of the causal force is closely linked to the problems of igneous activity.

Cause of Horizontal Compression. The outstanding question concerns the cause of folding and thrusting in the great mountain zones. When it was believed that the Earth consisted of a relatively thin crust resting on a hot molten interior, it seemed easy to explain buckling of the crust by assuming that there was steady contraction of the Earth's volume as the result of cooling. The already cold outer crust would be folded up as it gradually sank upon the shrinking interior, very much as the skin wrinkles upon an apple that contracts from drying. This view in its original simple form can no longer be held, because it is known from seismic evidence that the Earth is rigid, at least in the outer half of its radius.

Nevertheless a view commonly held to account for crustal deformation assumes contraction because of cooling. This is a survival of the idea mentioned above, but changed to accord more nearly with later knowledge. It assumes that the Earth, though rigid, is very hot within, and that progressive cooling causes slow shrinkage below a comparatively shallow depth. An English scientist has shown that this mechanism would account for a large amount of buckling in the outer shell, provided we make certain reasonable assumptions as to temperatures in the interior. Another form of the contraction hypothesis discards the idea of a hot interior, but assumes that the enormous pressures deep in the Earth cause slow but continuous increase in density and decrease in volume. The net effect would be the same as if cooling occurred, as the outer shell, unchanged in volume, would collapse on the shrinking core with consequent folding and thrusting.

It has been suggested also that vertical adjustments in the outer crust to maintain isostatic equilibrium (p. 17) may be a sufficient cause of folding and thrusting. This suggestion does not find support in evidences of enormous lateral pressure and displacement. In some mountain belts, individual rock sheets have been thrust horizontally for distances of 25 miles or more; and every large folded geosyncline represents lateral movement through tens of miles. It is difficult to conceive of a mechanism whereby slow vertical movements of the crust could be responsible for horizontal forces of such magnitude. The mechanism for maintaining isostasy must play an important role in some aspects of mountain development: first, in adjusting for loads when excess matter is crowded into a mountain zone by folding and thrusting; and, later, by restoring the balance disturbed when vast quantities of material are eroded from the high ranges (Fig. 6, p. 18). However, in this role isostatic adjustment can not be the principal actor. Unless there had been independent forces to upset equilibrium in the crust, all features of relief on the continents would long ago have disappeared through erosion.

Hypothesis of Continental Drift. A speculation which has won many advocates maintains that whole continents have shifted horizontally through long distances. It is claimed, for example, that Africa moved northward against the old Mediterranean geosyncline and crushed it to form the Alps and neighboring mountains; that the great chains of Asia were caused by southward shifting of that continent; and that the American cordilleras are the result of slow, long-continued westward drifting of North and South America. It is urged that no other explanation will suffice in view of the stupendous shortening recorded by mountain folds and thrusts. Students who favor the hypothesis of "continental drift" point out considerable geologic evidence which, in their view, strongly supports the concept. For example, if the maps of North and South America, as they appear on a globe, are moved eastward against Europe and Africa, not only do the continental margins match remarkably, but some old mountain belts in America—among them the Appalachians—appear to be continuous with mountain belts of the same geologic dates in lands east of the Atlantic. But what would furnish the motive power for breaking up and transporting continental masses?

One hypothesis that has received considerable attention attributes crustal deformation to slow-moving convection currents within a thick shell of the Earth. At first thought, it would appear that such currents are impossible, in view of abundant evidence proving rigidity in the Earth. It is urged, however, that a thick zone below the crust may be nearly devoid of strength, because of high temperature (p. 448). Some mathematical physicists agree that slow convection may operate in such a zone, provided an adequate source of heat exists. Advocates of the hypothesis assume that minute quantities of radioactive elements are contained in the rocks to depths of several hundred miles in the Earth, and that disintegration of these elements provides sufficient heat to set up convection. The supposed operation of such convection is as follows: A current rising below a large continental mass will be divided and turned laterally at the base of the strong crust; when the currents arrive at a continental margin, heat is lost through the ocean floor, and the cooled subcrustal matter sinks, thus completing the circulation. Frictional drag at the base of the crust may be strong enough to divide the continental mass and separate the parts. The forward edge of each continental fragment encounters resistance to movement and becomes deformed by folding and thrust faulting. Movement of the currents, and of continents propelled by them, is assumed to be at a rate imperceptibly slow. Separation of South America from Africa, through the 3000-mile width of the South Atlantic, is supposed to have begun more than 100 million years ago; hence the average rate of "drift" (assuming that it has continued to the present time) was less than two inches per year.

Although the hypothesis of moving continents seems fantastic to many students of the Earth, it has stimulated numerous investigations that are resulting in substantial increase of geologic information. Even if the hypothesis eventually should be disproved, its influence has been beneficial, and it may play an important part in revealing the secret of diastrophism.

SUMMARY

Mountains have been formed by differential erosion of plateaus; by volcanic accumulations; and by localized deformation of the crust. In the latter group we recognize several general classes of mountains based on types of structure, which reflect the action of diverse forces. The visible structure may be due to dislocation and tilting of crustal blocks, to simple doming of rocks, to folding with or without faulting, to large-scale thrust faulting, or to various combinations of these several processes. Connected with any type of movement there commonly have been injections and extrusions of igneous material, which have complicated the final structure of the mountain mass. Nearly all mountain-building movements have taken place in a series of pulses or phases distributed over a very long time; this is especially true of the chains that have complex structure. The cause of the great lateral pressure to which the belts of folding and thrusting bear eloquent testimony is an unsolved problem. We know from long study of many mountain belts that they were the sites of thick sedimentary accumulations during early stages of their development, and that deformation went on continuously or recurrently through long geologic periods.

Actual elevation to mountain heights commonly has followed the intensive folding and apparently was in large part independent of the deformation by lateral compression. Repeated uplifts have greatly prolonged efforts of erosion to destroy the great ranges. Many chains owe their present existence to differential erosion following broad upwarping of old folded and faulted belts. The continuous struggle between internal and external forces has left its clearest record in the diverse land forms of mountain regions.

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CHAPTER 20

LAND FORMS

GENESIS OF LAND FORMS

We have learned something of the various kinds of rocks that constitute the Earth's crust, we have examined the several processes that operate on the face of the Earth, and we have studied the intermittent movements of the crust that result from forces within the Earth. Having in mind these three factors, each of which exercises its own control over erosion and deposition, we are prepared to inquire into the shapes into which the land is sculptured. These are termed *land forms*, and their study is important not only because of their intrinsic interest, but also because an understanding of groups of these forms helps to reconstruct the later history of the part of the Earth in which they occur. The study of land forms is a means to an important end, as the following pages show.

Relation of Land Forms to Erosion and Deposition. The wave-cut cliff is a feature so distinctive that even where we find it well developed above present sealevel we infer that it was carved either by a lake or by the sea and was later affected by a change of level. Similarly the undercut sides and slipoff slopes of a stream valley are so characteristic of fluvial sculpture that, even where we find these two features facing each other across a valley with no stream between them, we infer that a stream formerly existed in the valley and that it carved the valley sides. Again, the discovery of a mountain crest partly consumed by the growth of cirques leads us to the firm conclusion that the crest was formerly sculptured by valley glaciers even though none now exist there.

The wave-cut cliff, the slipoff slope, the undercut valley side, and the cirque are land forms that are unusually distinctive. Each of these features characterizes the work of a specific group of processes: marine, fluvial, and glacial. Similarly the domes of the Yosemite (p. 38) are the work of weathering, the deflation hollows of the Gobi Desert and other arid regions (p. 202) are the work of the wind, the sinks (p.133) common in carbonate rocks are the work of subsurface water, and the fillings of lake basins (p. 147) are the work of lakes.

All but the last of the forms mentioned are erosional in origin. The same groups of processes, however, also build up depositional land forms. The (marine) beach and bars, the (fluvial) fan, the (glacial) end moraine, and the (eolian) dune are examples. The distinctive features of land forms, then, are determined by the processes that created them, and the processes include both erosion and deposition.

Relation of Land Forms to Rock Composition and Structure. Processes, however, are not the only factor involved in the making of land forms. A wave-cut cliff, an undercut valley side, and a cirque carved from resistant rocks are bolder, steeper, and more rugged than forms of the same types carved from weak rocks. Each is still recognizable; each still records the process that sculptured it; but each clearly indicates the influence of an additional factor—the general nature of the rocks from which it was carved. Furthermore, land forms also commonly reflect the structure of the component rocks. For example, the step-like character of the walls of the Grand Canyon (p. 111; Fig. 11) shows at once that the rocks are essentially horizontal. Similarly the appearance of the ridges in Fig. 316 shows that they are being carved from folded rocks.

Igneous activity and movements of the Earth's crust exercise the chief control over the composition and structure of many rock masses from which land forms are carved. For example, a volcanic cone evolves through a definite succession of land forms as it undergoes erosion by streams (p. 326). Fault blocks likewise go through a well-defined series of changes as erosion destroys their initial fault scarps and eventually creates fault-line scarps (p. 382).

Relation of Land Forms to the Geomorphic Cycle. Although rock character combines with erosion and deposition to control land forms, still a third factor is involved. The land forms between the streams in Fig. 64 are very different in appearance from those in Fig. 63 (p. 105), in spite of the fact that the processes at work and the rocks being worked upon are identical in both figures. The only difference between the two, indeed, is one of age, the land mass shown in Fig. 64 being farther advanced in the cycle than the one shown in Fig. 63. Age or position in the cycle, therefore, in addition to the rocks and the process at work, is an important factor in the shaping of land forms.

Value of Land-Form Study. Every land-form unit therefore contributes to the temporary aspect of a continuously changing landscape, for it is evolved from different features that preceded it, and it is certain to be altered into still other features by the processes that ceaselessly shape it. The practical value in the study of land forms lies in the fact that, because the character of any one unit usually betrays not only the general character of the rock composing it and the processes that sculptured it but also its position in the cycle, we can work backward from it and thus reconstruct landscapes that no longer exist. In this way we can read chapters in the history of the Earth that would be legible by no other means.

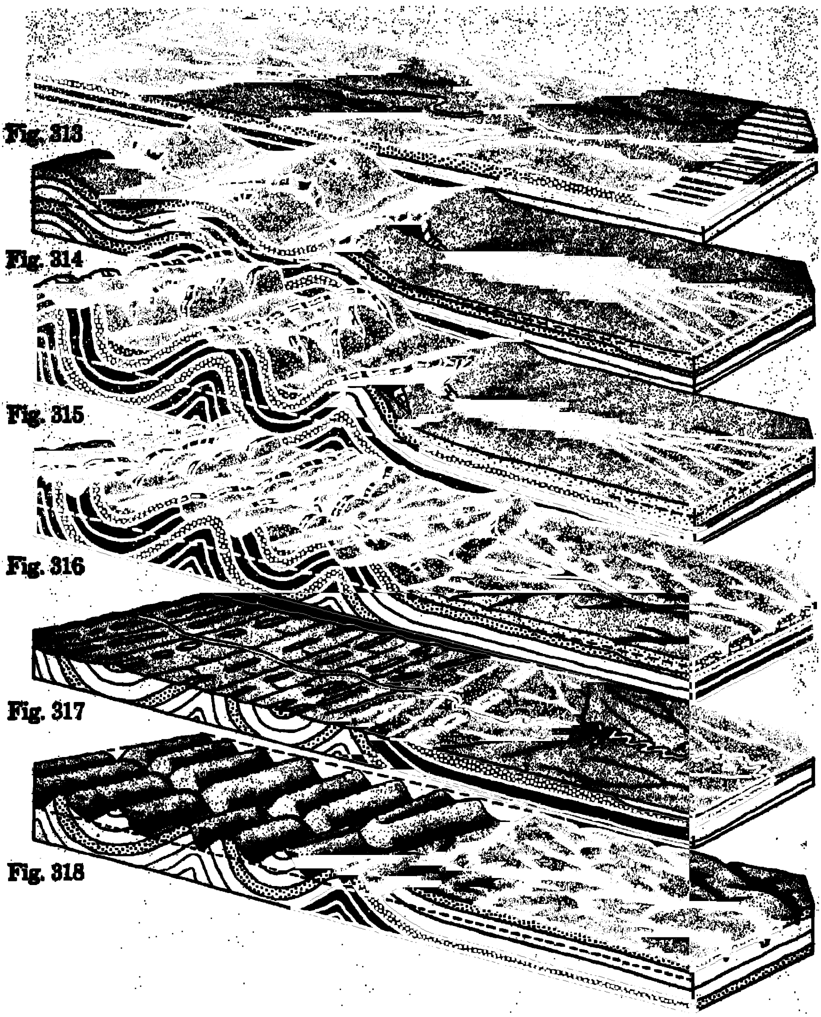
THE FLUVIAL CYCLE ON FOLDED STRATA

The folding and uplift of strata discussed in the preceding chapter are accompanied and followed by erosion, usually with streams and mass-wasting the chief agents. During the erosion of fold mountains there are evolved a number of characteristic land forms which are common in many parts of the world and are easily recognizable when their origin is understood. Therefore it is important to examine the fluvial cycle on folded strata in contrast with the fluvial cycle on homogeneous rocks.

In a moist climate the erosion of an area of folded strata is based on the same principles that control erosion of homogeneous rocks (p. 105). There are differences in detail, however, because of the unequal yielding to erosion of alternating weak and resistant rocks. Many variations are possible, but the following account, illustrated by Figs. 313–317, represents a typical cycle under these conditions. The illustrations are necessarily diagrammatic, in that fold structures change somewhat with depth, so that far below the surface they are likely to be less simple than those shown. However, all the land forms illustrated here are commonly developed, and the diagrams are geometrically correct.

Initial Stage. Let the scene be set with a land mass in old age (essentially a peneplane) developed on a series of horizontal sedimentary rocks, alternately resistant and weak (Fig. 313). The area is drained eastward by a main stream that reaches the sea on a broad delta. The tributaries, because the rock is horizontal, form an irregularly branching pattern (p. 99).

Early Youth; Antecedent Streams and Synclinal Tributaries. This section of the crust then begins to be compressed, and, as it gradually yields, it is very slowly bowed into a series of gentle folds (Fig. 314), of which one plunges northward. The folds are transverse to the course of the main stream, and the rate of bowing is so slow that it might be measured in inches per century.



Figs. 313–318. Fluvial cycle on folded strata in a moist climate. Greatest length of each block, approximately 20 miles. Vertical dimension greatly exaggerated.

Fig. 313. Initial stage, showing a land mass inherited from an earlier cycle.

Fig. 314. Stage of early youth, showing folding of the strata to form three anticlines; the main stream persists in its antecedent course and cuts water gaps, but the tributaries are obliged to follow new routes along the synclines.

Fig. 315. Stage of later youth, showing filling of valleys and breaching of anticlinal ridges, forming hogbacks.

Fig. 316. Stage of maturity, showing dissection of the earlier valley filling and growth of subsequent streams.

Fig. 317. Stage of old age, showing the weak rocks reduced nearly to baselevel, leaving monadnocks of resistant rock. Essentially a peneplane.

Fig. 318. The same area after rejuvenation by uplift, during which the streams have vigorously eroded the weak-rock areas, leaving the resistant-rock layers in ridges.

AERO SERVICE CORPORATION.

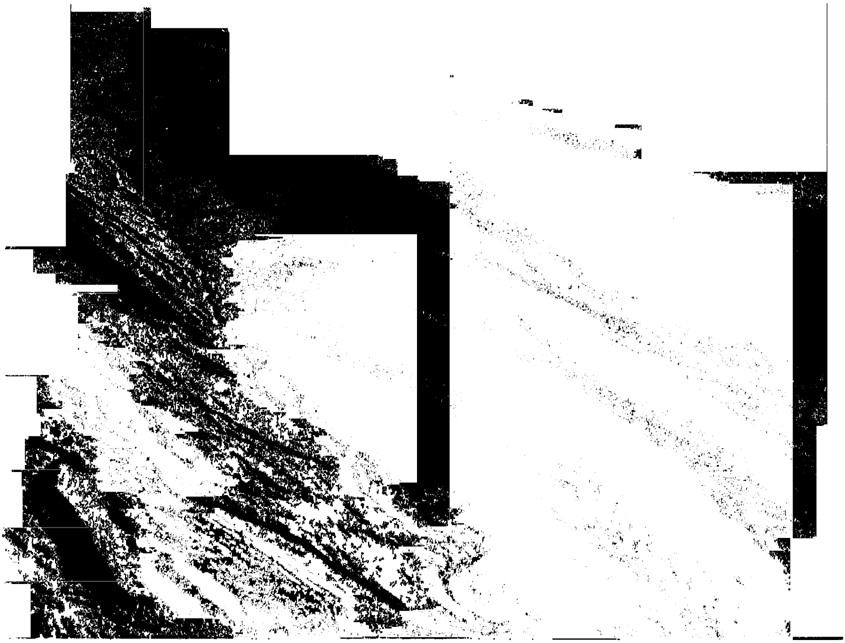
Fig. 319. Delaware Water Gap; view looking north (upstream). The Delaware River has cut the gap through Kittatinny Mountain, a sandstone ridge whose flat top is a remnant of a former peneplane. Pennsylvania lies to the left of the river, New Jersey to the right.

If the fold rose so rapidly that the stream could not cut through it, the stream might be *defeated*—converted into a lake by the damming effect of one of the anticlines, and forced to spill over or around the anticline at some new point. But, in the case selected, the main stream is just able to cut down through the slowly rising folds. A stream that is strong enough thus to maintain its original course in spite of folds or faults across its path is said to be *antecedent.* The Santa Ana and Temecula rivers, both of which cross ranges of hills southeast of Los Angeles, California, are antecedent to the uplifts that formed the ranges. The Columbia River and its tributary, the Yakima, are antecedent to anticlinal ridges in central Washington through which they have cut.

In the present example the tributaries, smaller and weaker than their main, do not fare so well; they are blocked off by the rising anticlines and diverted along the new drainage lines provided by the synclinal troughs. Concentration of drainage takes place in each of these, and pairs of tributaries are developed at right angles to the main stream, as consequents upon the folded surface. In this way the earlier stream pattern is extinguished and a right-angled pattern takes its place, with the chief tributaries occupying the synclinal troughs. At the same time, rapid runoff down the slopes of the anticlines forms steep consequent tributaries of the second order, which soon excavate deep gorges. As soon as these tributaries cut through the surface bed of resistant rock and penetrate the weak bed below, they cut the weak rock rapidly, undermining the resistant rock and quickly widening their valleys. This process is characteristic of the early stages of the cycle, in that the weak rocks are attacked as widely as possible, while the resistant rocks, escaping direct attack, fall piecemeal by undermining. The main stream crosses the anticlinal arches through *water gaps* (Fig. 319) which it has cut itself, and its increased load is deposited farther downstream. That the stream system is confronted by a long period of cutting is indicated by the volume of rock that lies above the new baselevel (indicated by a dashed line, Fig. 314).

Later Youth; Hogbacks. The folding has now reached its climax (Fig. 315), and the anticlines grow no higher. The crests thereby attain their maximum altitude above baselevel (dashed line, Fig. 315). But because much erosion has taken place since the uplift began, the actual mountains can never be as high as might be inferred at a later time by projecting upward the stumps of the anticlinal limbs. Steepening of the anticlinal slopes (compare Fig. 315 with Fig. 314) causes such greatly increased erosion by the tributaries of the second order that the streams in the synclinal troughs, whose gradients have not been steepened much since folding began, are not able to carry away the waste contributed to them. Therefore the troughs are partly filled with waste. The long ridges formed by the steeply dipping resistant strata are termed *hogbacks* (Fig. 320). They are common in the Colorado Rockies, the Big Horn Mountains, the Black Hills (p. 503), and many other anticlinal mountain masses. If the dip of the resistant bed were very gentle instead of steep, a more asymmetric and gently sloping ridge would result. A ridge of this type is called a *cuesta* (Figs. 335, 336).

Erosion of the weak layer, with undermining of the resistant layer above it, progresses most rapidly along the crests of the higher anticlines, because there gradients are steepest. Hence the anticline in the center (Fig. 315) has been breached along its entire crest, whereas the lower anticline to the east is only partially breached, and that only where its axis is highest. Because it is a plunging fold, its axis is inclined, and hence the tributaries of the second order that drain its



FAIRCHILD AERIAL SURVEYS.

Fig. 320 Hogbacks, at Rainbow Gardens, near Las Vegas, Nevada. Length of base of view, approximately 2 miles.

flanks have less steep gradients. This being the case, their cutting power is less than that of their neighbors on the other anticlines, and the resistant stratum that caps this anticline is therefore less readily breached. The structural skeleton of the whole region is now dissected into greatest relief, but the mass-wasting processes and the streams have much to accomplish before they can remove all the rock above baselevel and carry it into the sea.

Maturity. The weak strata continue to receive the most vigorous direct attack by the streams, and the resistant strata fall block by block as they are sapped and undermined (Fig. 316). Erosion of the weak strata in the middle anticline has now completely laid bare the rounded upper surface of the lower resistant stratum, and this likewise goes through the erosional process and is breached. Figures 315 and 316 show the appearance and development of other such rounded ridges in the hearts of the anticlines.

By this time the whole land surface has been reduced appreciably toward baselevel (dashed line, Fig. 316; compare Fig. 315), and the heights have been lowered sufficiently to decrease the supply of waste to the tributaries. The latter, freed from a part of their load, are again able to devote a part of their energy to downcutting, and they begin to dissect their previous deposits in the synclinal troughs and in the region east of the anticlines. By this time they have reached grade, and the hill slopes, continuously eaten away by mass-wasting, are also at grade.

Old Age. Although decreasingly effective, downcutting by the streams during late maturity and old age results in the reduction of the remaining highlands. The reason is simply that most of the weak rock has already been worn down close to baselevel, leaving little but resistant rock upon which mass-wasting can operate (Fig. 317; note that the dashed line has disappeared because the old-age surface practically coincides with it). Because of the great resistance of the only rock left above baselevel, and because gradients are slight, the process of erosion has become almost infinitely slow. Figure 317 depicts a peneplane with elongate monadnocks formed by outcropping edges of the resistant beds, together with one rounded ridge in the pitching anticline. Weak and resistant rocks alike are beveled by a surface of low relief. The cycle is virtually ended; from now on the streams can work only to destroy the monadnocks, with scarcely appreciable results.

Adjustment of Streams to Weak-Rock Belts. Figures 313–317 show that, although the initial tributaries develop as consequents in the synclinal troughs, they steadily decrease in importance until in old age (Fig. 317) they are almost completely extinguished. As the resistant rocks that form the crests of the anticlines are breached, tributaries develop in the underlying weak rocks and extend their valleys headward from the main stream along the exposed belts of weak rock. Streams that established themselves by growing headward along belts of weak rock in this manner are termed *subsequent* streams. As the cycle progresses, the subsequents, rapidly eroding their weak-rock valleys, steadily gain more drainage area at the expense of the consequents which have to cope with the resistant rocks in the synclinal troughs. In this way, through the process described on page 498, they acquire the lion's share of the runoff. From this the principle may be set up that *stream systems adjust their courses so as to flow as much as possible on weak rock and as little as possible on resistant rock.* Streams adjust their courses in other ways, but this is a very common method. Drainage developed in folded strata rarely reaches an adjusted condition before late maturity or old age. The more or less complete adjustment of the larger streams in the folded Appalachians is illustrated in Fig. 321. The strike of the folds can be readily inferred from the trend of the right-angled stream pattern.

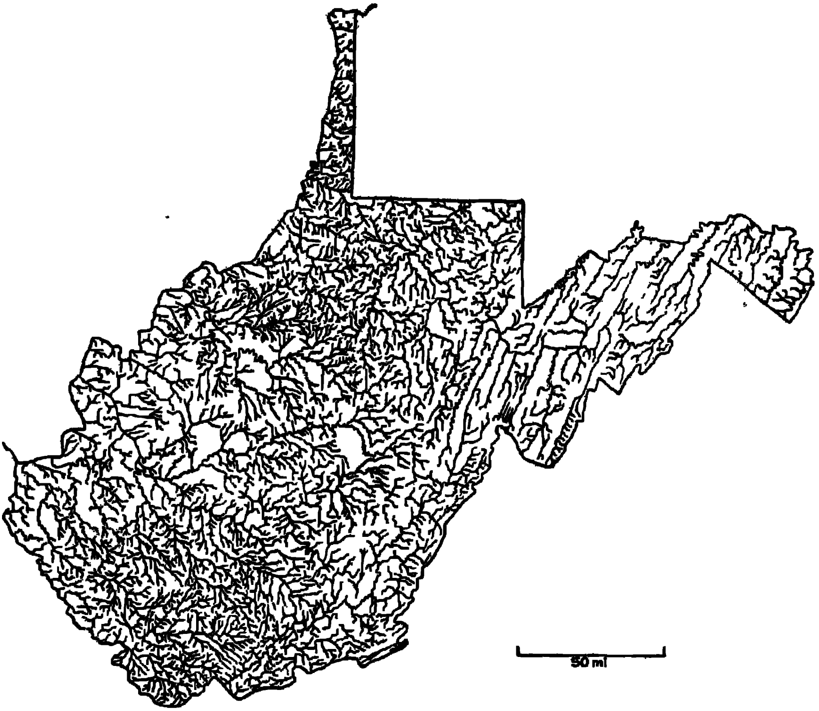


Fig. 321. Stream patterns in West Virginia. In the central and western regions the rocks are nearly horizontal and the pattern is irregularly branching. In the east the rocks are strongly folded along NE–SW axes and the resulting pattern is right angled, the long straight streams being subsequents which have cut their valleys headward in belts of weak rock. Note how thoroughly the region is drained (mature dissection in a moist climate), and compare Fig. 68, p. 109.

Uniform Beveling of Weak and Resistant Rocks. The significant fact emerging from the description of such a cycle is that, if left undisturbed, streams and mass-wasting wear down a land mass to *a peneplane in which both weak and resistant rocks are more or less evenly beveled.* Remnants of peneplanes exist in many parts of the world, each one cutting across rocks of various kinds and thereby testifying to the fact that it was formed under the close control of base-level, the only level at which streams and mass-wasting can cut down both weak and resistant rocks nearly to a common plane.

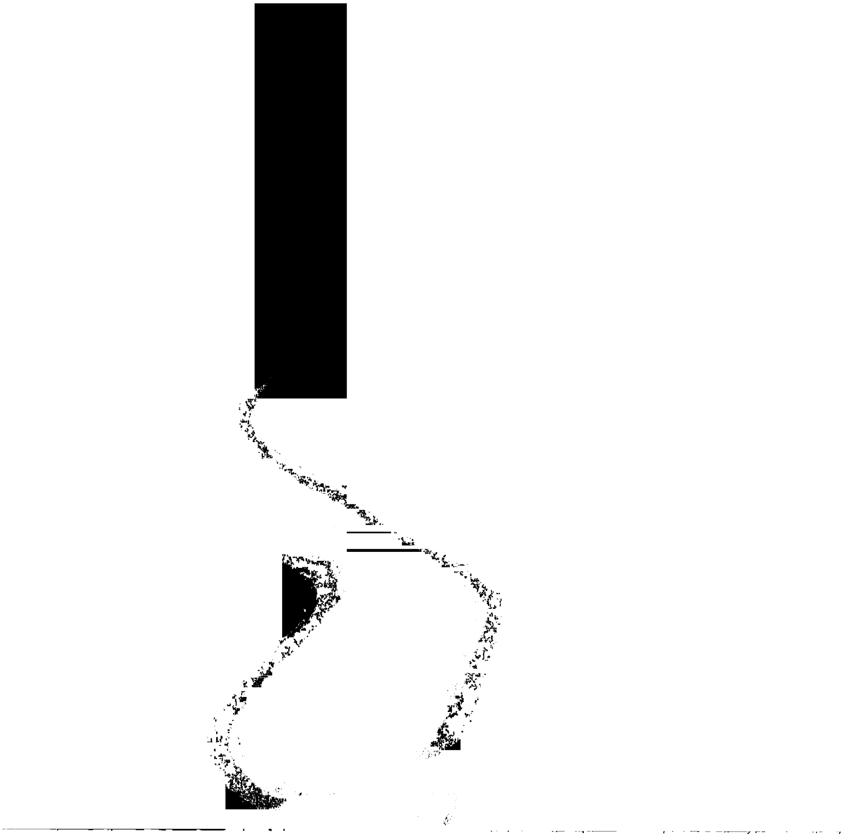
Peneplanes would be much more numerous if the Earth's crust were stable, allowing land masses to stand still throughout an entire cycle of erosion. What usually happens is that movement of the crust occurs long before the cycle is completed, either depressing the land, submerging the mouths of the valleys, and raising the baselevel, or causing the land to rise higher above the sea, increasing gradients and rates of erosion, and thereby starting the cycle over again. It is because of frequently repeated crustal uplifts of the land, forcing the streams to renew downcutting, that the continents have not been reduced to great low-lying peneplanes. Changes of level of land masses are therefore highly significant events in the history of the Earth. For this reason we must examine their effects on stream-sculptured land forms, so that by recognizing the sculptured forms in any region we can work back to the changes that produced them and thereby add still further to our knowledge of the Earth's past.

EFFECTS OF CHANGES OF LEVEL OF A LAND MASS

We have seen that the crust is subject to upward and downward movements of various kinds (p. 351) and that the level of the sea also fluctuates, in part independently of the crust (p. 195). If the evidence of change of level is distinct and widespread, the earlier cycle is said to have ended (even though it is not completed) and a following or *second* cycle begun. Thus the valleys shown in Fig. 323: *A, B,* and *C,* are now in a second cycle begun by change of level that has carried the land to a new position higher above baselevel.

The effect of relative downward movement of the land is most conspicuous along coasts, because there it consists of the submergence of valleys beneath the sea. The mouths of many of the larger valleys of the Atlantic Coast of North America are thus "drowned" (p. 239), as are many valleys in southwestern Britain and northwestern France.

Inland, downward movement of the crust is commonly accompanied by the accumulation of abnormally thick stream deposits. The sagging of basins such as the Salton basin (p. 152) may be accompanied by widespread and rapid filling with sediment. Downbending tends to reduce stream gradients on the floor of the downbent area, but at the same time it tends to increase gradients on the sides or flanks of the area, causing rejuvenation, as described below.



SPENCE AIR PHOTOS.

Fig. 322. Incised meanders, San Juan River west of Mexican Hat, Utah. The stream has been rejuvenated, incising the meanders to a depth of 1200 feet. Note the outcrops of horizontal resistant strata.

REJUVENATION

Movements of any kind that increase the slopes of the land increase the rates of erosion of the streams affected, by increasing their gradients. A stream given a markedly increased gradient by this means or any other is said to be *rejuvenated.*

Incised Meanders. One of the most conspicuous effects of rejuvenation is the incision of meanders. If the gradient of a meandering stream (p. 102) is increased so rapidly that the rate of lateral cutting by the stream can not keep pace with it, the stream is rejuvenated in its meandering course. The meander loops are gradually excavated into a continuous winding gorge whose bends are separated by partition-like, steep-walled projecting spurs (Fig. 322). These winding gorges are *incised meanders*, and are an indication that rejuvenation has taken place in the region where they occur.

The meanders can not remain deeply and narrowly incised for long. Lateral cutting is most rapid against those slopes that face up the valley, while the stream tends to slip away from the slopes that face down the valley (compare Fig. 56). By this process the projecting spurs are destroyed and a new strath is cut.

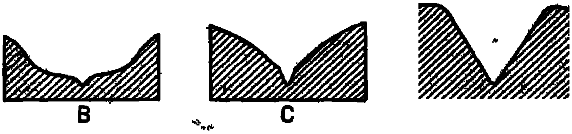


FIG. 323. Cross-profiles of valleys showing that the effect of rejuvenation depends on the stage at which it occurs.

A, old stream rejuvenated. effect very strongly marked. *B*, mature stream rejuvenated: effect strongly marked; former strath cut into terraces *C*, young stream rejuvenated: effect barely perceptible in the spurs along the valley sides. *D*, very young stream rejuvenated effect not registered. Compare Fig. 324.

On every continent there are more large streams with incised meanders than with normal meanders. This striking fact furnishes one more proof of the instability of the Earth's crust by recording the repeated upwarping of land masses and the rejuvenation of their streams.

Nongraded Tributaries. Downcutting by a rejuvenated main stream, whether meandering or not, induces rejuvenation of the tributaries by lowering their points of junction with the main. Rejuvenation works headward up each tributary valley, essentially as a new gully works up a slope. At any time during this process, the long profile of the tributary is steeper in its lower part than farther upstream. Many of the tributaries to the Merced River, which drains the Yosemite Valley in California, have strikingly steepened profiles as a result of rejuvenation of the Merced.

Valleys with Composite Cross-Profiles. A rejuvenated stream normally excavates a new V-valley within its former valley. If the latter is V-shaped, its cross-profile is scarcely altered (Fig. 323, *D*), but, if it had progressed into late youth prior to rejuvenation, the cross-profile of the new V-valley meets the profile of the old valley so as to form a pronounced shoulder (Figs. 323, *C*; 324). If the valley prior to rejuvenation was mature or old, the effect is very marked (Fig. 323, *A*, *B*). However, before such a composite cross-profile can

C. R. Longwell, U. S. Geological Survey.

FIG. 324. East end of Boulder Canyon, southern Nevada. Note abrupt changes in slope of valley walls, suggesting two rejuvenations. The three inferred successive cross-profiles of the canyon are indicated by dashed lines. Compare Fig. 323.

be safely ascribed to rejuvenation, it must be proved not to be controlled by differences of rock resistance to erosion (Fig. 70, p. 111).

Stream Terraces. A special type of composite cross-profile is represented by *stream terraces,* bench-like flats which are present in many valleys, and whose upper surfaces stand above the flood stages of the streams, thereby showing that they were not made under present regimens.

Geological Survey of Canada.

FIG. 325. Fill terraces along the Thompson River near Kamloops, British Columbia. The terraces are dissected by gullies.

Not uncommonly there are several terraces in series, rising away from the stream like two irregular flights of steps facing each other.

Generally they are small discontinuous remnants, but in some valleys they are continuous through long distances. Some (*fill terraces*, Fig. 325) consist entirely of stream deposits, the remnants of former thick valley fills. Others (*strath terraces*) are former straths cut into bedrock. Some terraces occur in pairs, each pair at a nearly uniform height above the stream (Fig. 326). The strath terraces are covered thinly with stream deposits laid down during the periods between the times of rejuvenation.

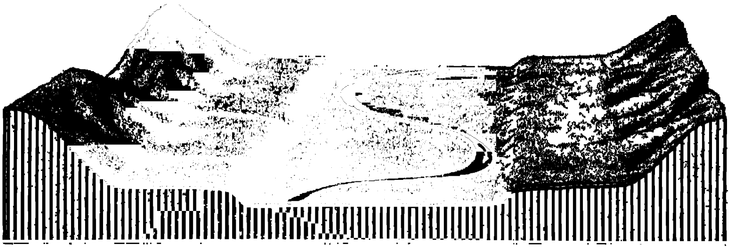


Fig. 326. Paired strath terraces caused by rejuvenation, since which the stream has cut a wide strath.

On the other hand, some valleys contain terraces that are not paired, that is, that do not "match up" at a common height on both sides of the valley. Ordinarily such terraces occur along a stream that is swinging from side to side of its valley as it slowly lowers its strath. At various points the stream encounters the bedrock that constitutes the valley sides. Deflected by these resistant obstacles, it swings away and is prevented from undercutting the fill that lies vertically above and immediately downstream from each outcrop of bedrock. The fill

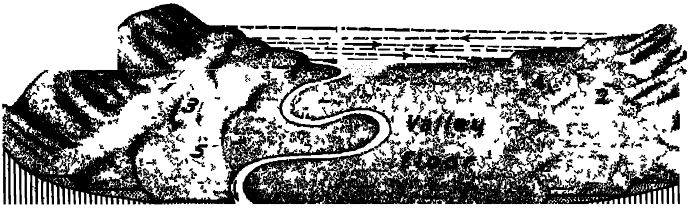


Fig. 327. Nonpaired fill terraces left by a downcutting stream that is meandering from side to side of its valley, encountering bedrock at various points. Dashed lines indicate the path followed by the meandering stream. The terraces are numbered 1 to 5 in the order in which they were cut. The present floor of the valley is trenched only by the stream channel.

remnants remain as terraces (Fig. 327), which gradually increase in height as the stream continues to cut downward. However, lateral migration of the channel takes place so very slowly that stream erosion lowers the channel somewhat between the time when it deserts one valley side and the later time when it encounters the opposite side. In consequence, no two terraces on opposite sides of the valley will lie at the same elevation above the stream.

Many nonpaired terraces have been formed by streams rejuvenated so gradually that they are able to swing widely as they cut downward; others may have been cut by streams which, having reached grade in an uninterrupted cycle (p. 81), slowly continued to lower their straths.

Remnants of Upbowed Peneplanes. It may happen that rejuvenation does not occur until the earlier cycle has reached the peneplane stage. The completion of the ideal cycle depicted in Figs. 313–317, is followed (Fig. 318) by rejuvenation during which the peneplaned surface (upper dashed line) is bowed up high above baselevel (lower dashed line). This permits the stream system to renew the work of excavation which had ceased under the old conditions. No new deformation of the rocks is required, for, although the surface itself has been smoothed out during the earlier cycle, the structure below the surface remains. The folded strata of unequal resistance, which disappeared as relief features through the beveling effect of peneplanation, are still there, needing only rejuvenation to bring them again into prominence. Hence a simple broad upbowing such as has occurred repeatedly throughout the Earth's history is all that is necessary. The sluggish streams are rejuvenated; they begin to cut actively, at the same time speeding up the process of mass-wasting over the whole surface. The weak rocks are again stripped away from the resistant rocks, which are left standing as prominent ridges. The tops of these ridges, which were beveled nearly flat during the earlier cycle, are wasted down so slowly and uniformly that during the early part of the second cycle they preserve the relative position of the former peneplane from which the intervening weak rocks have been cut away (Figs. 318, 328). The highest ridges of the Appalachian region are records of a former peneplane. Their beveled tops betray the fact that they are parts of a former continuous surface that stood close to baselevel, and that this surface was later dissected by erosion following rejuvenation (Fig. 319).

In Fig. 318 the weak rocks in the two synclines are temporarily preserved as ridges because they are protected by narrow caps of resistant rock. When these caps disappear by being undermined, the synclinal ridges will be quickly destroyed. The completion of the second cycle will see a new peneplane (developed in the plane of the lower dashed line). It will be much like that of Fig 317 but will differ from it in at least two respects: (1) Monadnocks will be fewer because of the complete removal of the upper resistant layer from the two synclines. (2) The remaining monadnocks will be farther apart because the resistant layers will be intersected farther down the limbs of the anticlines by the plane of the new baselevel. These ridges, apparently fixed in position, are in reality slowly migrating as they are eroded down the dip of the resistant strata of which they are composed

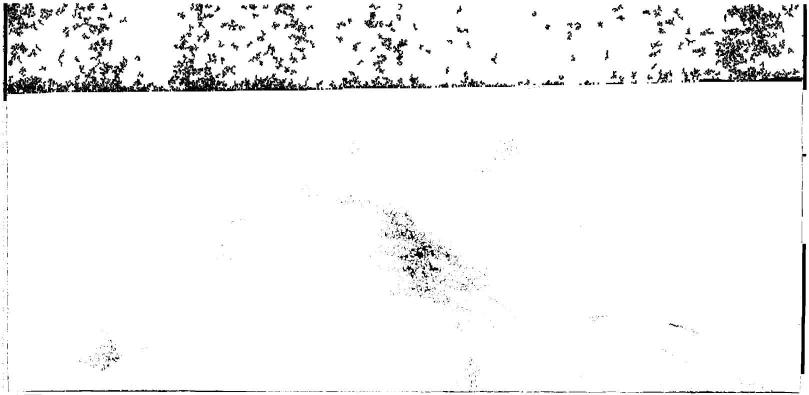
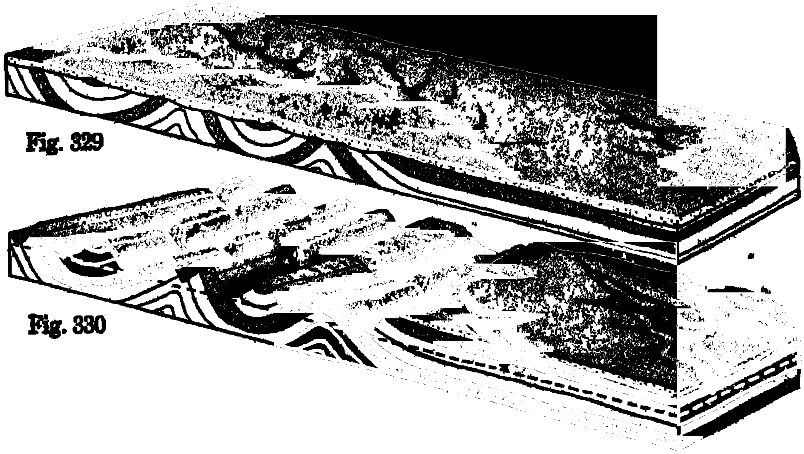


Fig 328. Peneplaned mass arched up to about 900 feet above sealevel and dissected to a stage of maturity in a cycle that began with the uplift. The tops of the hills and ridges together form a remarkably even skyline They are remnants of a peneplane that beveled several different kinds of rocks, all of them strongly folded Restigouche River near Campbellton, New Brunswick

Superposed Streams. Many large streams cut across the strike of weak and resistant rocks without any regard to their structure. Such a discordant relationship, we have seen, may be brought about by crustal bending or folding across the path of a previously existing stream, in which case the stream falls into the class of antecedent streams. In Figs. 314–316 the antecedent character of the main stream is shown by remnants of the folded original land surface still preserved in some places. In Fig. 318, however, every vestige of the original folded surface has been destroyed, and consequently at this stage there is no means of proving the antecedent origin of the stream. But it is clear that the tops of the ridges record a former peneplane, and that the main stream, formerly flowing on the peneplane, was incised in its present course through rejuvenation of the peneplaned land mass.

An extensively eroded land area may be buried beneath later deposits, so that these rest upon the underlying structure with angular unconformity (p. 387). Blankets of this kind have been formed in many regions by fluvial, glacial, volcanic, lake, and marine deposits.



FIGS. 329–330. Superposition of a stream.

FIG. 329. The peneplaned land mass of Fig. 317 has been submerged, covered unconformably with horizontal sedimentary strata, and re-elevated, and new streams have formed on the uparched surface.

G. 330. Later stage, after the streams of Fig. 329 have cut down through the unconformity into the discordant rocks beneath. The horizontal strata have been removed by on except along the tops of the resistant ridges.

In time the streams that develop on the surface of such a blanket cut through it into the very different rocks and structure below, upon which they are said to be *superposed* (Figs. 329–330). The stream shown in Fig. 196 is superposed upon a denuded laccolith. Many of the large transverse streams in the Rocky Mountain region such as the Platte River system have been superposed from thick and extensive covers of alluvium.

Small remnants of the overlying cover remain through part of the cycle as caps on the hills. As long as they remain, they testify to the manner in which the streams acquired their discordant courses, but after they have been stripped away it may be very difficult to determine whether the nonadjusted streams, cutting water gaps through ridges of resistant rock, were antecedent to the structures they cross, were let down upon them through the rejuvenation of a peneplaned region, or were later superposed upon them. Discrimination is important because it may involve wide differences in the dates we infer for the origin of the nonadjusted streams. In either case, we find in both kinds of streams still another evidence of the instability of the crust, and of the complexity of sculpture of the lands to which crustal movements give rise.

Stream Capture; Wind Gaps. A water gap abandoned by the stream that cut it, and therefore left as a streamless notch in a ridge, is a *wind gap* (Fig. 331). Water gaps become wind gaps commonly as the result of capture of the gap-cutting streams by subsequent streams which, occupying belts of weak rock, can erode very rapidly. This process is illustrated by Snickers Gap in the Blue Ridge of Virginia 15 miles south of Harpers Ferry. The floor of this wind gap hangs 700 feet above the Shenandoah River at the west base of the ridge. A small stream, Beaverdam Creek, heads at the east base of the ridge and flows east. During an earlier cycle the land around the base of the Blue Ridge stood as high as the floor of Snickers Gap, which contained a transverse stream, the ancestral Beaverdam Creek (Fig. 332). But the much larger transverse ancestral Potomac River, which crossed the ridge 15 miles farther north, was able to cut its gap downward much more rapidly into the resistant rocks that formed the ridge. The Shenandoah, a subsequent stream tributary to the Potomac, was

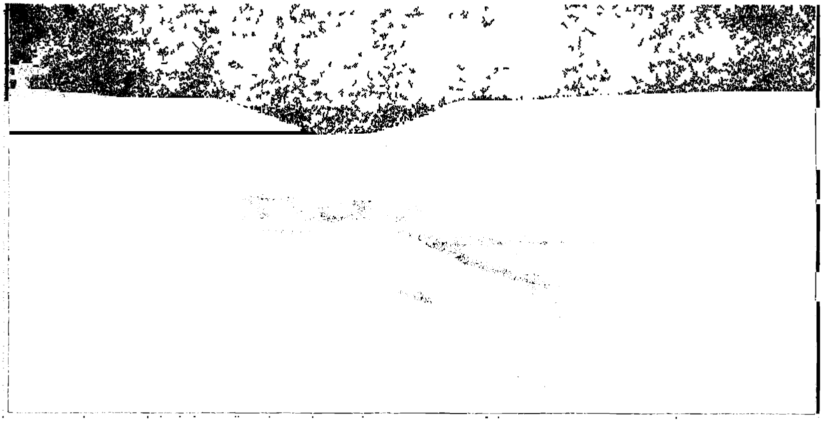
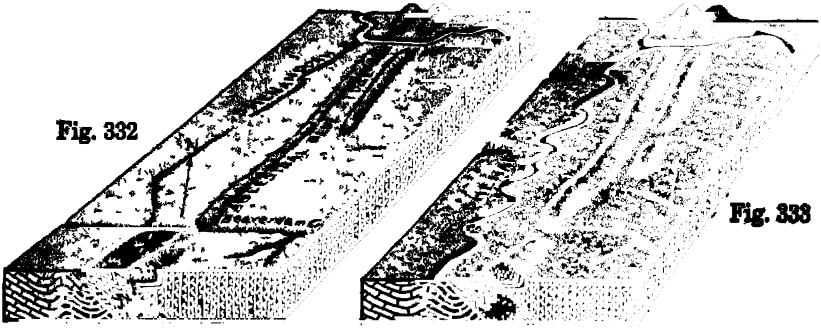


FIG. 331. Dolls Gap, West Virginia, a wind gap in New Creek Mountain, an anticlinal ridge of resistant sandstone. The gaps shown in Fig. 319 and Fig. 331 are related to each other in the manner shown in Fig. 334.

thereby given a relatively steep gradient, and so lengthened its valley headward toward the south, wore down the divide that separated it from Beaverdam Creek, and captured the latter by diverting its water toward the Potomac (Fig. 333). Thus in consequence the gap was abandoned, and the shortened creek (the present Beaverdam Creek) was confined to the area east of the Blue Ridge. As the Shenandoah and the small streams east of the ridge continued to cut downward, the ridge was left standing in greater relief, with the abandoned gap high and dry.

Some wind gaps may have been formed by relatively weak streams which, affected by folds and faults athwart their courses, were able to maintain their original courses only for a time and were then forced to abandon them. On the other hand, some ridges have notches in them that were made by small streams and mass-wasting processes localized along transverse joints, faults, and other structures. These must be carefully distinguished from wind gaps.



FIGS. 332–333. Conversion of a water gap into a wind gap by stream capture. Long dimension of block about 25 miles. Vertical dimension exaggerated.

FIG. 332. Former drainage across the Blue Ridge in northern Virginia.
FIG. 333. Present drainage resulting from beheading of Beaverdam Creek by the subsequent Shenandoah. Note increase in height of the ridge as the weak-rock areas adjacent to it are wasted away by erosion.

Stream capture like that outlined above occurs repeatedly during the adjustment of a large stream system to belts of weak rock. Capture occurs also in several other ways, two of which are explained on pages 105 and 115.

HISTORY OF COMPLEX LAND FORMS

Thus far we have examined land forms sculptured by some one process only, in simple rocks of simple structure during one or at most two cycles. In many regions land forms have a more complex origin, but still they can be understood by applying to their study the same principles used in interpreting simpler forms. We can best understand these complex features by seeing actual examples, and can begin with those characteristic of the Appalachians.

HISTORY OF THE APPALACHIANS

The central part of the Appalachian region as we see it today consists chiefly of long parallel ridges formed by the outcrops of resistant rocks, such as sandstones and conglomerates, in a strongly folded sedimentary series (Fig. 334). Most of the ridges, though narrow, have even summits, broken here and there by water gaps (*WrG*) and wind gaps (*WdG*). The tops of the ridges reach a common plane (*1*, Fig. 334), whereas the lowlands (*2*, Fig. 334) between the ridges are a network of small hills whose summits are nearly accordant. They are drained by small streams which flow into larger meandering subsequent streams (*SSS*, Fig. 334) following the strike of the strata, and all the meanders are incised. The subsequents in turn are tributary to great streams that flow southeast indiscriminately across resistant-rock ridges and weak-rock valleys. These streams are the Delaware, the Susquehanna, and the Potomac.

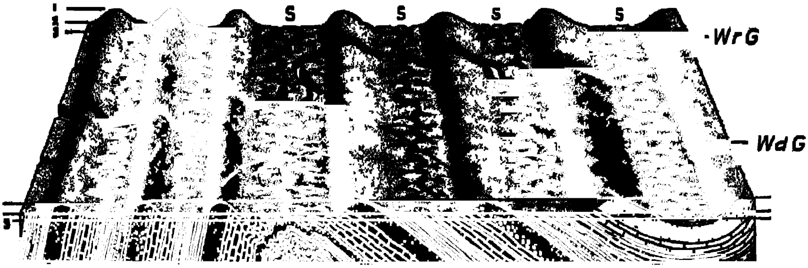


FIG. 334. Ideal segment of a part of the central folded Appalachian region. The main stream is nonadjusted, trenching seven ridges of resistant rock through seven water gaps (*WrG*). A former stream course, long abandoned, is indicated by a wind gap (*WdG*). The present tributaries are subsequents (*SS*). Control by three successive baselevels (*1*, *2*, *3*) is indicated by accordant ridge tops, an accordant group of hilltops in the weak-rock belts, and the present stream profiles. Compare Fig. 318.

With these facts in mind, how much of the Appalachians' erosional history can we reconstruct? *First*, a period of strong deformation, as revealed by the great folds whose eroded stumps appear in the ridges. Erosion must have begun during the folding and have continued throughout at least one long cycle, resulting at length in a peneplane. This we know because the resistant rocks in the folded series have been beveled down nearly to a common plane. Of course these rocks did not exist as high ridges during the peneplane stage, but were merely gentle swells on a low gently undulating surface near sealevel. *Second,* slow upbowing of the whole region, beginning a second cycle by rejuvenating the streams, most of which, being already in weak-rock courses, excavated the weak rocks again, leaving the resistant strata once more projecting as ridges. The gap (*WdG*) must have been converted from a water gap into a wind gap early in this second cycle, because its floor lies between the altitudes of surfaces *1* and *2*. Abandonment of the gap occurred probably through capture of the gap-cutting stream by a weak-rock subsequent (*S;* see also Figs. 332, 333). The new surface (*2*) was of more restricted extent because of the great quantity of resistant rock remaining unreduced in the ridges. *Third,* another upbowing, rejuvenating the streams and causing them to incise their meanders and with the aid of their tributaries to dissect their former valley floors (*2*) into the present network of small hills.

Events of this kind, reconstructed through study of land forms, have occurred in the Appalachian region. Because geologists are not agreed as to the number of rejuvenations that have affected this region, the number shown here is not of essential importance. The underlying principle of successive rejuvenations is unaffected by the number of upwarpings that may have occurred.

The main transverse streams (one of which is shown in Fig. 319) are not adjusted to the structure and have clearly been in this condition since before the date of the peneplane (*1,* Fig. 334). How they originally acquired their transverse courses has not yet been fully established. Neither is there agreement as to the date at which each surface was formed, but the whole sequence of events from the folding of the strata down to the present required about 200 million years.

It is evident that, in a region like that described, the highest ridge tops record a widespread peneplane, whereas the tops of the low hills record that at a later time only the weak-rock areas were reduced to form a surface close to baselevel. The difference between the higher peneplane and the lower "uncompleted peneplane" is only one of degree. Where the degree of perfection and the extent of a former surface can not be inferred from the still-existing remnants of it, the convenient though somewhat ambiguous general term *erosion surface* is frequently applied.

HISTORY OF THE PARIS BASIN

In northern France lies a region composed of alternating weak and resistant layers of marine sedimentary rock bent down into a broad shallow syncline and considerably eroded. The outcropping edges of the resistant strata give the whole basin the appearance of a low stack of dishes, each dish smaller than the one underneath it (Fig. 335).

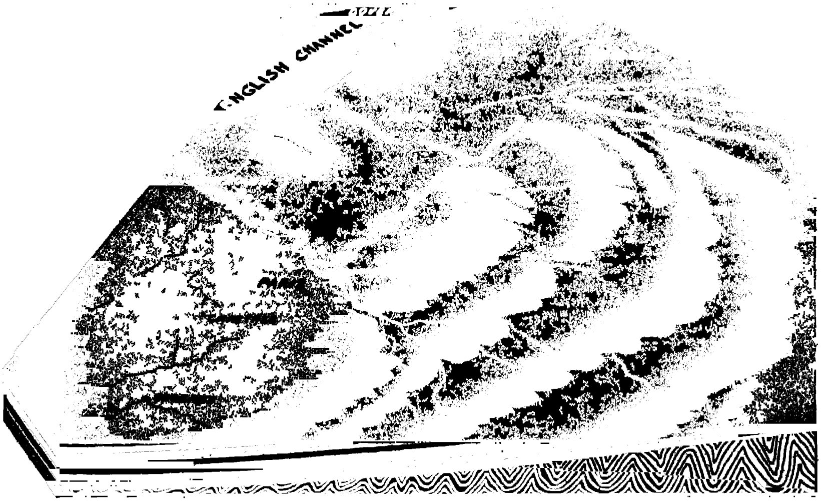


FIG. 335. The Paris Basin, developed in a shallow syncline. Each of the outcropping resistant strata forms a prominent cuesta. The River Seine is shown with its partly adjusted tributaries. Length of front of block 200 miles.

Paris lies in the uppermost dish, which derives its title of *Ile de France* from the fact that its rim was once believed to be a wave-cut cliff marking the shore of a former island. The drainage pattern, like that of the Appalachians, consists of main streams (the Seine system) out of adjustment with the symmetric structure, and tributary streams some of which are adjusted to the weak-rock belts beyond the edges of the resistant strata.

Upon first examination one might infer that the edges of the dishes, like the synclinal troughs in Fig. 314, were the immediate consequence of the downbending. But the edges are beveled, like the far higher and more resistant ridges in the Appalachian region: they are therefore remnants of a former peneplane and are not first-cycle forms but second-cycle forms. The history of the region was somewhat as follows:

After the sedimentary rocks were laid down in a former shallow sea, they were gently bent so as to form a broad syncline and were lifted above sealevel. Erosion at length reduced the entire area to a peneplane, beveling weak and resistant rocks alike. On this peneplaned surface the major streams followed the nonadjusted courses they have at present, although how they acquired these courses is not certain.

The peneplane was then bent upward, and the streams were rejuvenated, causing the headward growth of a new generation of valleys along the weak-rock belts, forming a right-angled drainage pattern. Excavation of these valleys left the resistant rocks standing in relief as cuestas, which differ from hogbacks (p. 486) only in that they are formed by gently dipping instead of steeply dipping strata.

The growth of subsequent streams has resulted in partial adjustment of the streams to the weak-rock belts; but the major streams still pursue the nonadjusted courses they inherited from the peneplane recorded by the beveled tops of the cuestas.

The concentric arrangement of cuestas brought about by the basinlike structure gives the French capital a unique series of natural fortifications. Furthermore, because of variations in the dip and character of the strata, cuestas are more strongly developed on the east side of the basin than on the west. This arrangement was of great advantage to the Allied armies on the Western Front during World War I, because it presented a series of difficult natural obstacles to the armies of the Central Powers, advancing from the east.

HISTORY OF THE BLACK HILLS

The Black Hills in southwestern South Dakota (Fig. 336) are a mountain mass 100 miles long and 50 miles wide. The highest peaks stand 4000 feet above the surrounding territory. The center consists of a core of ancient metamorphic and igneous rocks locally cut by much younger igneous intrusions. Belts of sedimentary rocks chiefly of marine origin cut by the same intrusions are arranged symmetrically around it, dipping away from the central mass and neatly framing it. The frame of sedimentary rocks is made more distinct by the fact that the upturned edges of the more resistant strata form hogbacks and cuestas whose scarps face the central mass. The lowlands between the cuestas are drained by subsequent streams, which in turn are tributary to larger streams that radiate outward in all directions, cutting water gaps through the cuestas. The hogbacks and all but the highest ridges and divides on the central mass are beveled by an erosion surface resembling an imperfect peneplane. Here and there remnants of a once

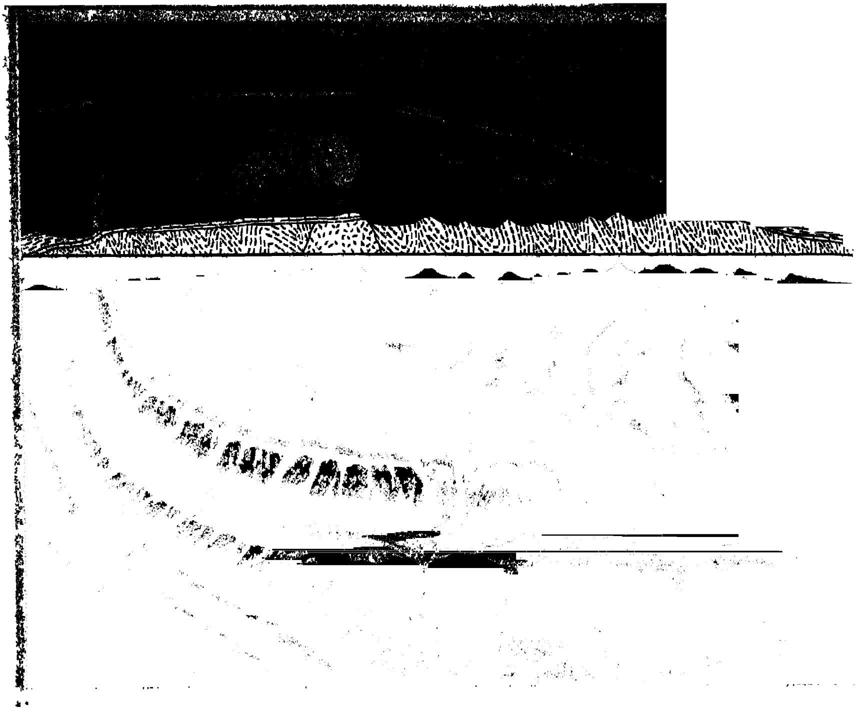


Fig. 336. Perspective view of the Black Hills, looking north The drawing has been pulled apart to show, in much generalized form, the geologic structure along a west-east line through its middle part. Length of geologic section about 80 miles. The Hills are seen to be developed on a large dome, breached, and surrounded by hogbacks and cuestas made by outcropping resistant sedimentary strata.

continuous veneer of stream-deposited conglomerate, sandstone, and shale lie unconformably upon it.

From these data we can reconstruct the chief events in the history of the Black Hills. In shallow seas that formerly covered the entire area, the marine sedimentary strata were built up on the surface of the ancient rocks. The whole mass was then arched up into the form of a great dome. The arching was accompanied by intrusions of magma, some of which formed small laccoliths, making subsidiary domes in the sedimentary rocks here and there.

Streams developed on the great dome, flowing off down the slopes and forming a pattern like the spokes of a wheel. The dome was breached (Fig. 315). Subsequent tributaries developed and cut their valleys headward along the strike of the weak layers of sedimentary rocks, leaving the resistant layers standing out as hogbacks and cuestas, each of which migrated steadily outward down the dip of the strata as erosion continued, thus laying bare more widely the central core.

The cycle begun by the updoming resulted at length in the reduction of the area to an incomplete erosion surface truncating the dome, and leaving a central group of monadnocks composed of the ancient basement rocks. While the erosion surface was being cut, the streams deposited gravel and sand upon it, shifting their courses and gradually building up a blanket of fluvial sediment that covered the surface.

A second updoming rejuvenated these streams, and as they cut down once more they became superposed from the unconformable cover onto the more complex structure below. If we disregard minor rejuvenations recorded by strath terraces, we may say that the streams are still in the cycle started by this uplift, and that they have thus far only partially adjusted themselves to the weak-rock outcrops on the surface on which they were superposed. Bits of the erosion surface are preserved at the tops of many of the divides, some of them still capped by ragged remnants of its unconformable cover.

The history of the Rocky Mountains in Colorado and Wyoming is similar in many respects to the history of the Black Hills. The evolution of the Rockies has been complicated, however, by the introduction in the higher ranges of a cycle of valley glaciation (p. 188) now in its waning stage Glacial sculpture has imparted a serrate skyline to the high Rockies which is lacking in the nonglaciated Black Hills. There is evidence that the evolution of both the Rockies and the Black Hills from their folding to their present condition has required some 60 million years.

DISSECTED PLATEAUS; THE GRAND CANYON REGION

The land forms of the Grand Canyon region, although sculptured from structures much simpler than those of the Appalachians, the Paris Basin, and the Black Hills, are more impressive because of their great size The initial land mass which is being carved up is a great plateau that extends over much of Utah, Arizona, New Mexico, and western Colorado (Fig. 289, p. 451). The underlying rocks consist of both marine and nonmarine sedimentary strata that have been *lifted* thousands of feet with little deformation. Here and there the nearly flat-lying beds are interrupted by faults and flexures of minor significance compared with the plateau mass as a whole.

The surface of the plateau in the vicinity of the Grand Canyon is a well-developed erosion surface that bevels the minor irregularities in the strata. Into this peneplaned surface the Colorado has incised its canyon (Fig. 290), with a course not adjusted to the structure represented by the flexures and faults. On the peneplane the stream evidently followed this course, in which it persisted when the surface was bent broadly upward to form the plateau. The cause of the discordant course upon the peneplane is not certainly known. The uplift of the peneplane to a height of 6000 to 8000 feet above sealevel began the present "canyon cycle" of erosion, during which the rejuvenated Colorado has cut a gash locally more than a mile deep and 220 miles long. In spite of the tremendous depth of excavation, the canyon cycle has not yet progressed beyond the stage of youth. The tributaries have not worked far headward from the main stream and have succeeded in making but little impression on the vast plateau surface. Near the river, where dissection is greatest, divides between the tributaries have been reduced to narrow spurs, and some of the spurs have been carved up into huge strings of pyramids by mass-wasting graded to the tributaries (Fig. 290). Many of the pyramids are of mountainous size and would certainly be described as mountains if their origin as detached remnants of the near-by plateau were not so obvious.

The excavation of the canyon after the uplift of the peneplaned mass may not have required much more than a million years. After a far greater time has elapsed, however, the tributaries will have cut their valleys so far headward as to carve the plateau into a network of ridges and residual remnants, destroying the identity of the plateau and converting it into an intricate maze of buttes and spurs of mountainous size.

This process gives us a key to the origin of many masses commonly described as mountains, such as the Catskill Mountains, the Allegheny Mountains, and the mountains of central Idaho. In the first two the rocks are essentially horizontal strata, whereas the rocks of the Idaho mountains are chiefly massive granite. In all three regions, however, the summits lie near a common plane. It is evident that such masses as these are merely dissected plateaus (p. 453).

HISTORY OF THE GREAT LAKES REGION

The Appalachian region, the Paris Basin, the Black Hills, and the plateaus of the Grand Canyon region show us that two or more cycles of erosion have had a hand in the modeling of many of the land forms of today. In each of the four regions described, the cycles were brought to an end by rejuvenation, permitting the same process, fluvial erosion, to operate on them anew. The Great Lakes region, on the contrary, has had a different history. The region as a whole is a great plain whose surface consists chiefly of glacial drift left by the invasion of broad ice sheets (p. 191). The plain is interrupted by the basins of the several Great Lakes. The drift (Fig. 130, p. 187), although very thick in some places, is so thin elsewhere that it fails to mask the topography of the bedrock beneath it.

From a study of this topography, supplemented by data gained from the records of thousands of well-borings, we can reconstruct the landscape as it was before the ice sheets flowed slowly down from the north and covered it. The elongate basins now occupied by the lakes were then large stream valleys belonging to one or more well-developed systems fed by an intricate network of tributaries. The topography, in a stage of late youth or early maturity, may have resembled that of central Kentucky and Tennessee today. If glaciation had not occurred this region would have continued to evolve through the cycle on which it had begun.

But the advent of the ice sheets introduced a second (glacial) cycle, during which the former large stream valleys were converted into lake basins by a combination of glacial erosion, local warping of the crust, and the heaping up of glacial deposits to form dams. When the ice sheets melted they left the surface so profoundly altered that a third (fluvial) cycle was instituted. The principal valleys of the former stream system were filled with glacial deposits, and many of them were entirely obliterated. The runoff was forced to follow courses consequent upon the newly made surface, substituting a new drainage pattern for the old.

The Great Lakes region therefore affords an example of a three-cycle development of a group of land forms in which the second and third cycles were introduced not by rejuvenation but by *change of process.* The glacial (second) cycle has left an impress upon the present (third) cycle that can not be obliterated for a long time to come, because even though the glacial deposits should be completely removed by erosion, the streams developed on their surface would inevitably be superposed on the underlying bedrock, with which they would be out of adjustment. The former presence of the unconformable cover could therefore be inferred even though it had been entirely stripped away.

CONCLUSION

The foregoing discussion brings out the fact that land forms are the response of the surface of the land to various processes acting for various lengths of time upon various kinds of rocks having various structures. Alter one of these factors and the resulting land forms will be different. The process factor depends chiefly on climate. Thus a change from a moist climate to a dry climate may cause the development of interior drainage, change the dominant type of mass-wasting from soil creep to talus creep, and increase the efficacy of the wind as an agent of erosion, whereas a change from a temperate climate to a cold climate may freeze the water into perennial ice and bring on a cycle of glaciation.

The general character of the rocks likewise is of vital importance to the land forms modeled by any of the erosional processes. For example, the landscape sculptured from a mass of weak rocks elevated only slightly above the sea can never acquire bold relief even in maturity, because gradients are small at the outset. On the other hand, resistant rocks lifted high above sealevel will be cut by valleys with steep gradients and steep sidewalls, which yield very slowly to the changes wrought by erosion during the progress of the cycle.

Thus it appears that the seemingly unending variety of landscape is controlled in reality by comparatively few factors, and that variations in these factors result in the differences in scenery that lend enjoyment to travel and add to the value of human existence.

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A most interesting account of the war of 1914–1918 on both western and southeastern fronts, laying special stress on the influence of land forms on the character and conduct of the fighting.

10. Geology from Original Sources; by W. M. Agar, R. F. Flint, and C. R. Longwell. Chapter 12, Land Forms. Henry Holt & Co., 1929.

Selections from various authors, on land forms in various parts of the world.

11. Geologic History of the Yosemite Valley; by F. E. Matthes. 137 pages. U. S. Geological Survey, Professional Paper 160, 1930.

Explains the formation of the valley by stream action and glaciation, and outlines the late-geologic history of the Sierra Nevada.

12. Geomorphology of the North Flank of the Uinta Mountains; by W. H. Bradley. U. S. Geological Survey, Professional Paper 185, pp. 163–204, 1936.

A detailed and critical presentation of the later geologic history of a classic region.

CHAPTER 21

MINERAL RESOURCES

RELATION OF MINERAL RESOURCES TO GEOLOGY

Modern civilization is strikingly dependent on mineral products. The mineral industry is second in importance only to agriculture. The world's mineral resources were more heavily drawn upon in the first quarter of the present century than in all previous history. The resources are being consumed at an unprecedented rate. Practically, this draft on our mineral resources means that the known deposits are being rapidly depleted and that in order to maintain the high rate of production new deposits must be discovered fast enough to keep pace with exhaustion. The depletion of our domestic mineral resources was greatly accelerated by the enormous demands of World War II.

The easily found mineral deposits of economic value have already been discovered, so that more ingenuity and more work are necessary to find new deposits. Therefore the cost of discovering new deposits is steadily rising. It is the counsel of prudence, then, to enlist all possible aids to minimize the financial risks of exploration. The application of the principles of geology to the search for valuable mineral deposits is one of the main functions of economic geology. Powerful auxiliaries to aid in this search have been developed, notably the seismograph for "seismic prospecting," the torsion balance, and other geophysical instruments. We call these instruments auxiliaries advisedly, because the results obtained by means of them must be interpreted in the light of the geology of the area examined before they can be practically used in finding valuable mineral deposits.

More important even than the aid given by applied geology in finding new deposits is the power it gives to guide intelligently the development and exploitation of the deposits already found.

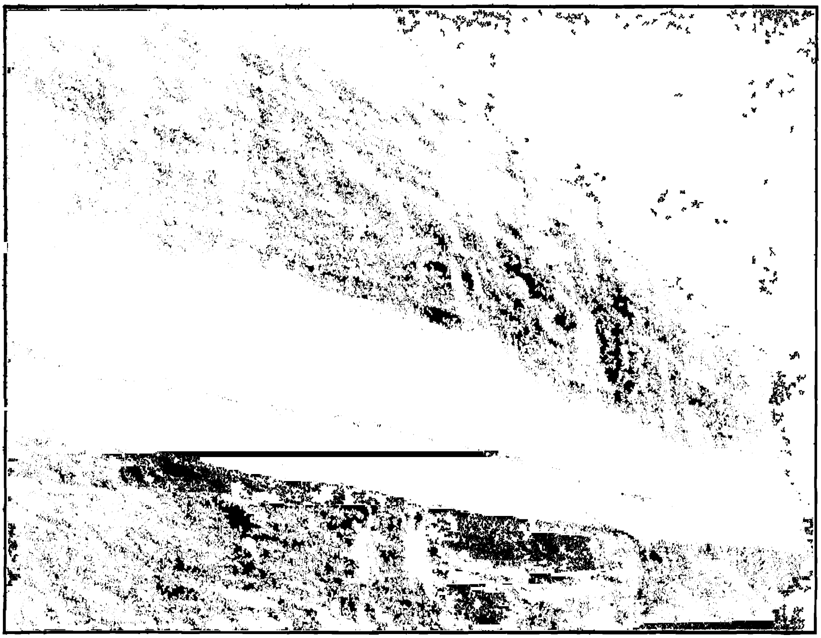
A mineral deposit of economic value is a concentrated body of substance that originally was thinly scattered through parts of the envelopes of the Earth. A coal bed is made up of carbon that has been concentrated by the photosynthetic power of plants from the carbon dioxide thinly diffused throughout the atmosphere; an oil pool is an accumulation of petroleum generated from carbonaceous matter thinly scattered through certain sedimentary strata; an iron-ore deposit is a concentrate of iron that was disseminated through the Earth's crust. The problem of how valuable mineral deposits have been formed is in essence, then, the problem of how the scattered material was assembled. It is mainly from this point of view that the following account is written. All the major geologic processes—including igneous action, sedimentation, metamorphism, weathering, and stream transport—have under certain circumstances brought about valuable concentration of minerals.

COAL

Coal is the chief mineral fuel. The per capita consumption in the United States, having risen from 2.5 tons per year in 1890, has recently shown a tendency to become stabilized at 5 tons per year. Although coal is a far more prosaic resource than gold or the industrial metals, it vastly outdistances them in value. The Pittsburgh coal bed is the most valuable single mineral deposit in the world. In Pennsylvania alone it has already yielded $6,850,000,000. Nearly 4000 million tons of coal have been mined from it in Pennsylvania (without including West Virginia, Ohio, and Maryland), and in 1948 it still contained in Pennsylvania 6800 million tons of recoverable coal.

Occurrence and Nature of Coal. Coal is a compact mass of carbonized plant debris. Peat, the embryonic form of coal, has already been described as to mode of origin (p. 155). It ranges from a brown to yellowish matted mass of interlaced fibrous vegetal matter, greatly resembling compressed tobacco, in the upper part of the accumulating bed, to a dark brown or black homogeneous mass, much resembling clay when wet, in the deeper, lower part. Between peat and the most mature coal, anthracite, there exists a complete series of gradational varieties of coal. This gradation led to the conclusion that coal, no matter how structureless and amorphous it may appear, consists of carbonized plant remains. This conclusion has now been fully confirmed by the results of modern microscopic study, which show that coal is not amorphous but consists of plant debris in various stages of alteration. Fragments of wood, bark, leaves, roots, spores, seedcoats, and lumps of resin can be recognized in coal. Even in such highly carbonized material as anthracite the plant ingredients can be made visible and, unexpectedly enough, are found to be so well preserved that the species of many of the plants can be identified.

Coal occurs as beds, which as a rule are inclosed between sandstones and shales (Fig. 337). The occurrence of coal in strata of this kind harmonizes with other evidence that the vegetal matter from which the coal was formed accumulated in swamps. Limestones are rare, either immediately above or below a coal bed. Where they do occur, they indicate that the swamp was near the sea and at sealevel and therefore likely to be inundated by the temporary incursions of the sea During such an incursion the coal-forming debris becomes buried under marine sediments. This has happened many times during the making of the British coal fields. Many important coal fields, however, such as the Saar, extending from Lorraine eastward into Germany, were formed in inland basins.



M R Campbell, U S. Geological Survey

FIG. 337. **Coal bed, 8 feet thick, containing two clay partings. Glendive, Montana**

Most coal beds are underlain by a layer of clay, the *underclay*, which, because it contains the roots of plants, is interpreted as a fossil soil. Stumps of large trees, whose root systems penetrate the underclay, are abundant in some coal fields. These features support the prevailing idea that most coal beds have been formed from plant matter that accumulated where it grew.

Thickness of Coal Beds. Coal beds range from hardly more than a carbonaceous film to hundreds of feet in thickness. Some are of small extent, but others persist over thousands of square miles. The bituminous coal beds in Pennsylvania (about 60) range up to 100 feet in thickness; but most of them are from 1 to 4 feet thick. About ten are extensively mined. The anthracite beds on the average are much thicker, the thickest bed being 114 feet, of which 105 feet is clear coal.

In Germany some of the beds of brown coal are more than 300 feet thick.

Chemical Composition of Coals. The composition of a coal is customarily ascertained in determining its value as a fuel, and in the ordinary analysis the following constituents are reported: (1) fixed carbon, the carbon that is left after the volatile matter has been driven off; (2) volatile matter, mainly combustible hydrocarbons ("gas") but including some inert gases, such as carbon dioxide; (3) moisture; (4) ash; and (5) sulphur. The heating value of a coal depends on its content of fixed carbon and volatile combustible matter. Moisture, ash, and sulphur are undesirable ingredients.

Coals containing much ash are said to be *low grade.* The better-grade coals carry less than 10 per cent of ash.

Classification of Coals by Ranks. Coals are classified according to *ranks* on the basis of their fixed-carbon content and their physical properties. Four groups of coals are recognized, in order of increasing rank as follows: I, lignitic; II, subbituminous; III, bituminous, and IV, anthracitic. There is a continuous gradation from the lowest-rank to the highest-rank coal; consequently the divisions between them are necessarily arbitrary.

The lowest-rank coals are the *lignites,* so called in reference to their obviously woody appearance, or *brown coals,* in reference to their color. These immature coals when taken from the mine may appear to be perfectly dry, yet they contain 30 to 40 per cent of water. On exposure they lose most of this water; they slack and crumble to pieces and become dangerously subject to spontaneous ignition.

Subbituminous coals constitute the next higher group. Their black color and lack of woody structure distinguish them from lignites, and their tendency to slack distinguishes them from the bituminous coals. The next higher-rank coals are the *bituminous.* They do not slack on exposure. They generally have a layered or "banded" structure and show a cross-jointing—a prismatic jointing perpendicular to the banding. The low-volatile members of this group comprise the coals of highest calorific value, surpassing even the anthracites.

The highest-rank coal is *anthracite,* distinguished physically from bituminous coal by its conchoidal fracture and absence of cross-jointing. Water and volatile matter are extremely low, and nearly all the carbon is fixed carbon. Anthracite because of its low volatile content is smokeless during combustion; and this feature together with its cleanliness makes anthracite the ideal coal for domestic use.

The increase in the rank of a coal, then, is marked by progressively larger content of fixed carbon and by decreased content of water and volatile matter.

Relation between Rank and Geologic History of Coal. Coal has been formed in all the geologic periods since the Devonian, when a flora first became established on the lands. Most coal was formed in forested swamps. The requisite conditions for a coal bed to form are at least two: (1) the swamp must be stagnant, hence poorly ventilated, thus preventing the plant debris from oxidizing to carbon dioxide; and (2) the swamp must be slowly sinking, thus permitting progressive accumulation of the plant remains.

The nature of the environment in which a coal bed was formed is inferred from the botanical character of the coal-forming flora and from the character of the inclosing sedimentary strata. For example, the Eocene brown coals of Germany were formed under tropical conditions, as inferred from the presence of rubber-bearing trees, whereas the younger brown coals of Germany, since they contain much coniferous wood (Sequoia), evidently accumulated under conditions differing greatly from those of the older coals. Modern analogues of coal-forming swamps are the warm-temperate Dismal Swamp of Virginia and North Carolina, a cypress swamp originally 2200 square miles in extent; the subtropical Everglades of Florida of 10,000 square miles; and the dense mangrove coastal swamps of tropical Sumatra and northeastern Borneo.

Plant matter changes into coal in two stages, a biochemical and a geochemical. The biochemical stage consists of a bacterial fermentation (p. 155), which sooner or later is arrested by the toxic products formed during the bacterial activity. The substances of which the plants are composed, a dozen or so, are destroyed during this fermentation in a definite order: the protoplasm, the most sensitive, disappearing first, and the waxes and resins, the most resistant, last. Accordingly, the duration and intensity of the biochemical stage determines how much of the various plant substances becomes the ingredient material of the future coal. The composition thus acquired influences somewhat the later history of the coal.

The later changes in the coal are termed geochemical, because they are brought about by geologic activities. Time was regarded formerly as an important factor in causing coal to advance in rank. However, one of the oldest coals, near Moscow in Russia, some 300 million years old, is still in the condition of lignite. Much evidence indicates that coal matures rapidly, geologically considered. In nearly all coal fields the deeper coals contain less volatile matter and more fixed carbon than the upper beds. This relationship is known as Hilt's law, but the rate of increase in fixed-carbon content differs in different fields. It is probably a geothermal effect, as previously pointed out in Chapter 17. Folding involving internal deformation within the coal bed is apparently necessary to bring about advances to the higher ranks. Anthracite generally occurs in strata that have been closely folded. However, coal beds have been acutely folded without change in rank, probably because of insufficient confining pressure to cause metamorphism. Under severe deformation, as in Rhode Island and the French and Swiss Alps, the coal tends to become graphitic and hence incombustible.

Coal, then, is sensitive to metamorphism; in fact, it is far more sensitive than other rocks in the Earth's crust and gives evidence of metamorphism long before the associated strata show any effects. By means of chemical analyses the changes in the coals brought about by metamorphism can be determined accurately; and, as many analyses of coals are available in most coal fields, regional variations in the intensity of metamorphism can be shown on a map.

Where coal and oil occur in the same area, practical use is made of coal as an indicator of metamorphism in the search for oil. Experience has shown that, in those tracts where the coal has been devolatilized so that the fixed-carbon content exceeds 65 per cent, oil is not likely to be found, and exploration in the direction of increased fixed-carbon content is therefore unprofitable.

Coal Reserves of the Nation. The extent of our unmined coal reserve is a matter of great national interest. In computing the amount of this reserve several assumptions must be made—and should be clearly pointed out so that the estimate will have definite meaning. Some of these assumptions are (1) the extreme practicable depth of mining, here taken to be 3000 feet; (2) the minimum thickness of the coal beds that can be included in the estimates; (3) maximum ash content permissible—30 per cent is the figure adopted, which admittedly includes very low-grade coal. The national reserve of coal is estimated to be 3500 billion tons. Some 700 billion tons more are estimated to lie below the surface from 3000 feet to 6000 feet. The staggering total, 4200 billion tons, will last for several thousand years, on the assumption that the present rate of consumption will continue. The higher-rank coals, especially those close to industrial centers, are likely to be exhausted, however, in the foreseeable future.

At the present time the coal reserve is regarded as constituting our most likely substitute reserve for petroleum when our petroleum reserves are exhausted. Coal can be liquefied to yield a crude petroleum, and this can be distilled to yield gasoline. According to present knowledge the coals most suitable for this processing—hydrogenation—are those of subbituminous rank (800 billion tons, equivalent to 800 billion barrels of gasoline), practically all of which are in Wyoming, Montana, and the Dakotas.

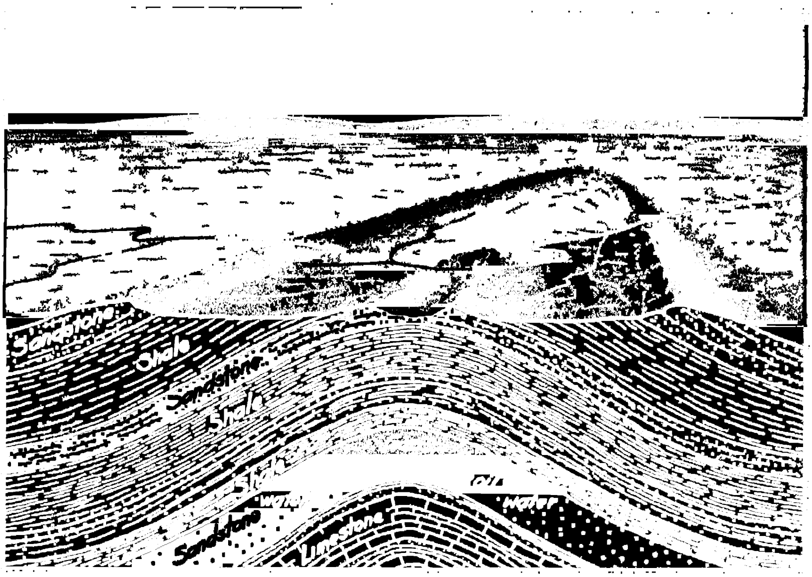
OIL AND GAS

Oil, or petroleum as it is more formally called, consists of gaseous, liquid, and solid hydrocarbons in mutual solution. As there are many of these hydrocarbons (compounds of carbon with hydrogen) and as they are mixed in various proportions, the oils differ markedly in the various oil fields. A distinction is commonly made between *light oils*, those much lighter than water, and *heavy oils*, which are near water in specific gravity. The light oils have a higher content of gasoline than the heavy oils. As the gasoline content influences the commercial value, the light oils command a higher price per barrel than the heavy oils.

Requisite Conditions for the Occurrence of Oil. Four conditions must be fulfilled in order that oil can accumulate in the rocks. First a *source rock* must be present which contains the necessary carbonaceous matter from which the oil can be formed. Apparently it is necessary that this carbonaceous matter should consist of plant remains of a special kind high in hydrogen, such as spores, which thus supply the hydrocarbons of which petroleum is composed. As a result of experience it appears that marine bituminous shales are the most common source rocks; limestones are rarely source rocks. Why rocks of fresh-water origin should not give rise to valuable accumulations of oil has not yet been ascertained, but the fact of their unfavorable nature is of great practical interest, for it excludes from consideration areas underlain by such strata because they are most unlikely to contain valuable accumulations of oil.

The second requisite is a *reservoir rock* in which the oil can accumulate. Most commonly the reservoir rock consists of sandstone, whose pores are large enough to allow the oil to move freely through it. Less commonly limestone and dolomite serve as reservoir rocks. In sandstones the average pore space is 15 per cent, and the range is from 5 to 40 per cent; in other words, a sandstone of maximum porosity can hold eight times as much oil as one of minimum porosity.

The third requisite is a favorable structural arrangement of the strata, which is the condition that determines the place where the oil collects in the reservoir rock. The simplest form of such a favorable geologic structure is an anticline. It was the first structure that was recognized as controlling the accumulation of oil, but now nearly twenty kinds of structural traps are known to occur in the world's oil fields.



Modified from Hewett and Lupton, U. S. Geological Survey.

FIG. 338. Gas, oil, and water as they occur in an anticline. Gas, being the lightest, occupies the crest of the arch; oil, being heavier, lies below the gas; and water, the heaviest of all, is at the bottom.

The fourth requisite is an impervious layer, generally shale or clay, which overlies the reservoir stratum, and which with the structural arrangement forms a structural trap in which the oil is held.

All four conditions are fundamental for the occurrence of oil; without any one of them commercial accumulations of oil are impossible. Inasmuch, however, as structure controls the actual underground position of an oil accumulation, and therefore in geologic practice the structure is determined as accurately as possible in advance of putting down a well, the structural requisite is of major practical importance. The mode of occurrence of oil in an anticline is shown in Fig. 338.

If, because of insufficient pressure, free gas occurs with the oil (under heavy pressure all the gas would be dissolved in the oil), there will be this triple arrangement: gas fills the pore space in the reservoir rock in the arch of the anticline; below it oil fills the pore space in the reservoir rock; and below this occurs water, generally so charged with salts as to be termed a brine. The fluids are thus arranged according to their specific gravities: the brine, which is heaviest, is at the bottom; and the gas, which is lightest, is at the top. The oil that fills the pores in the body of porous rock forms what in popular parlance is called the "oil pool."

If the sedimentary strata contain no water, the oil instead of occupying the crests of the anticlines occurs in the bottoms of the synclines. Numerous favorable traps have resulted also from movements on faults that have brought impervious against pervious strata.

Recognition of the requisite conditions governing the accumulation of oil makes it possible to block out in a given territory those areas that do not contain oil and those that possibly contain oil. Areas of igneous and metamorphic rocks, except under most exceptional conditions, contain no petroleum, and areas of nonmarine strata, although they may contain oil, are in general highly unfavorable. Finally it must be pointed out that, although areas likely to contain oil can be found by the skilful use of geology, with or without geophysical aid, the actual presence of oil can be proved only by putting down a drill hole and tapping the oil.

Stratigraphic Traps. In 1930 the East Texas field was discovered, the greatest oil field ever found. Up to January 1, 1948, it had yielded 2480 million barrels of oil, and the recoverable oil still in the ground was estimated to be 2500 million barrels. The magnitude of the East Texas field can be appreciated when we consider that an oil field is regarded as a major field if its total yield is 20 million barrels. The conditions that governed the accumulation of the oil in the East Texas field differ notably from those in structural traps. The oil occurs in the wedge ends of sandstone layers that are truncated at an acute angle by an unconformity; it was trapped in the upslope wedge ends of the sandstone layers where they are covered by impermeable beds of the overlying unconformable formation (Fig. 339). Such an oil-bearing reservoir is termed a *stratigraphic trap*.

It is believed that the future trend in oil-finding technique will be to develop the ability to find more stratigraphic traps, for the more obvious structural traps have mostly been discovered.

Life of an Oil Well. The gas, oil, and salt water in an oil-bearing structure, such as shown in Fig. 338, have in the course of geologic time attained an equilibrium. As soon as the top of the structure is perforated by the drill this equilibrium is violently disturbed, and the disturbance spreads radially from the well. At first the well produces gas; then it becomes a gusher or flowing oil well; later it has to be pumped; and finally salt water appears with the oil, presaging the extinction of the well. In many oil fields withdrawal of the gas causes the oil to move to the upper part of the structure, and the water follows up after the oil.

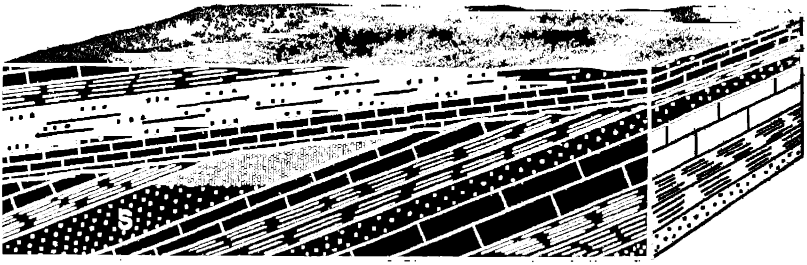
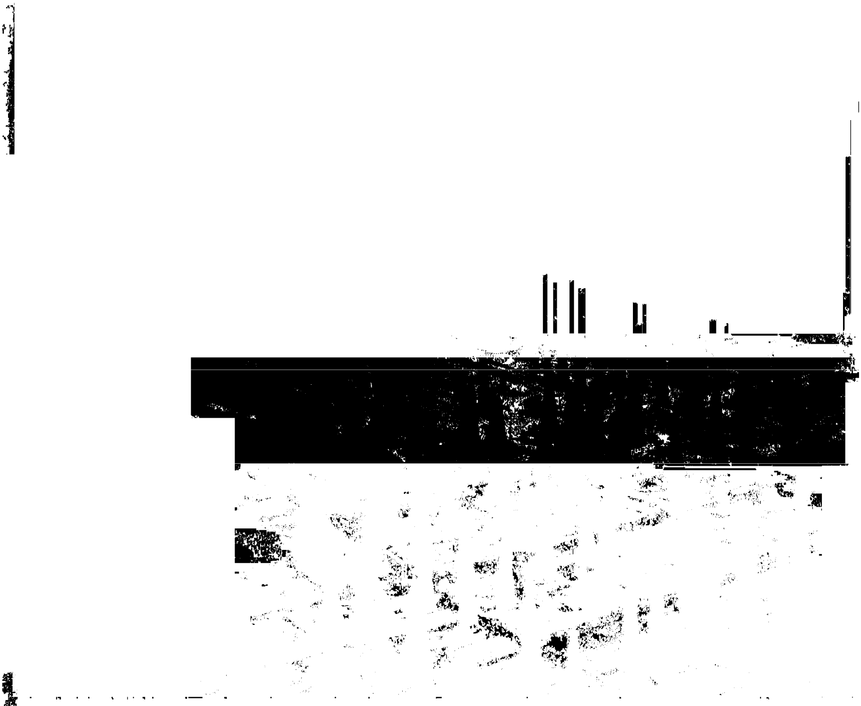


FIG. 339. Stratigraphic trap, showing oil (shaded pattern) occurring in the wedge end of a sandstone bed (*S*) below an angular unconformity. Length of block is about 3 miles.

Oil, being a fluid, moves readily in the direction of decreased pressure. Consequently it recognizes no property lines, and if the pool is under divided ownership the first wells that tap the reservoir drain not only the oil within their own tracts but also, if situated near the property lines, the oil from the adjacent tracts. If the owner does not take out his own oil, his neighbors will. This fact accounts largely for the close spacing of wells in certain fields (Fig 340) and for the wasteful overproduction that ensues soon after the discovery of a new field.

Coarsely porous rocks yield their oil freely and give rise to extremely productive gushers (Fig. 341). But the wells become exhausted rapidly—a fast life and a short one go together here. Finely porous rocks hold on to their oil tenaciously and so prolong the life of a well into old age. The record-breaking Mexican wells, the most spectacular of which yielded as much as 260,000 barrels a day, drew their oil from limestones, which like most limestones contained solution channels and cavities. The oil therefore escaped with enormous rapidity and gave



SPENCE AIR PHOTOS.

Fig. 340. Signal Hill oil field, southern California.

the wells their astounding yields, and for the same reason their period of production was short.

Function of Gas in Oil-Field Operation. The chief gas associated with oil is methane, CH_4, the simplest of all the hydrocarbons. This gas and other allied gases, collectively called "the gas," are dissolved in the oil under heavy pressure. The great importance of this natural gas in the efficient operation of an oil field is now fully appreciated. In most fields it is the chief agent that brings the oil to the surface: it drives the oil to the well and lifts it to the surface; it reduces the viscosity of the oil and the tenacity with which the oil is held in the pores of the rocks. This tenacity is so great that after an oil field is "exhausted," that is, after it ceases to yield oil by pumping, most of the oil, generally as much as 75 per cent or even 85 per cent, is still underground. Less oil is left in the ground if the gas content is properly utilized. The more efficiently the gas is utilized the smaller the number of cubic feet of gas that is required to bring a barrel of oil to the surface.

Because the gas has these important functions and because its withdrawal from any part of an oil pool eventually affects pressure conditions throughout the whole pool, some states have made it unlawful to waste the gas content of an oil pool. An oil pool is a natural unit, and to operate it most efficiently and to obtain the maximum recovery of oil it must be operated as a unit.

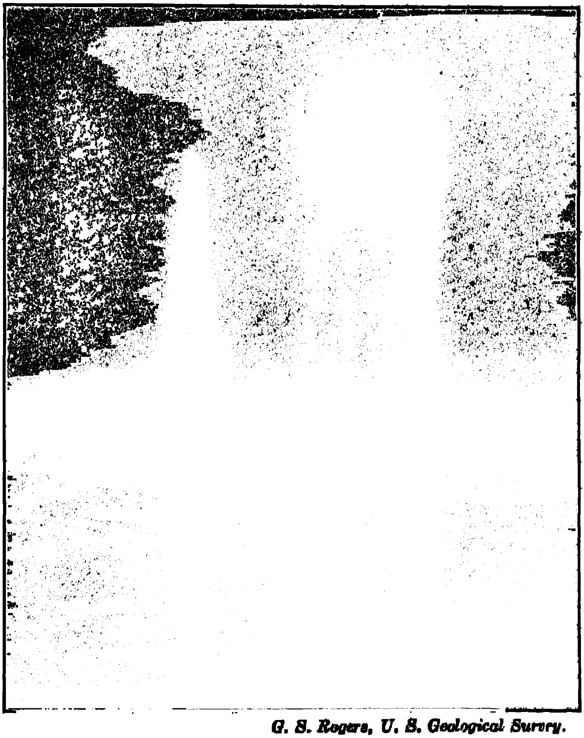

G. S. Rogers, U. S. Geological Survey.

FIG. 341. Gusher in Sunset oil field, California, shortly after oil was struck.

It is common practice in some fields to pump back into the ground, through wells otherwise not in use, gas that comes to the surface unavoidably in the recovery of oil. This operation is called repressuring.

Geophysical Prospecting. One of the most remarkable of the seven major oil regions of the United States is the Gulf Coast field of Louisiana and Texas. The oil is associated with cylindrical or steeply conical masses of rock salt which, forced to flow plastically under heavy pressure, have punched their way up from unknown depths through the overlying sedimentary strata. These masses are termed *plugs,* and most of them are a mile or so in diameter (Fig. 342). The tops of some of the salt plugs are practically at the surface, but many of them are deep below the surface. Oil occurs in association with salt plugs in three kinds of traps: (1) cap rock, a peculiar rock made of calcite, gypsum, and anhydrite occurring as a capping over the tops of the salt plugs; (2) flanking sands, abutting upon and cut off by the salt plug; and (3) supercap sands, the sandy strata that arch over the tops of the plugs, forming structural domes. Few of the salt-plug fields, however, produce oil from all three structures.

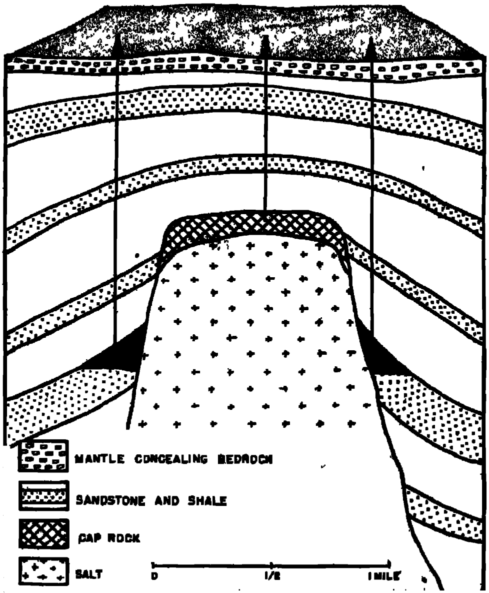


FIG. 342. Salt plug and enveloping beds, showing the occurrence of oil (black shaded pattern) in the flanking beds.

The first of these oil fields to be brought in was the famous Spindletop, near Beaumont, Texas, discovered in 1901. Spindletop is a circular mound rising 65 feet above the otherwise level monotony of the Gulf coastal plain. The top of the salt plug is at a depth of 1100 feet, and the mound at the surface appears to have been due to a renewed upward push by the salt below. Many salt plugs, however, are without mounds or any other indication at the surface that they are present at depth.

In recent years, geophysical methods have been used with brilliant success in locating these hidden salt plugs. In essence the various geophysical methods depend on the fact that the physical properties of the salt differ greatly from those of the surrounding rocks. The seismograph (p. 404) began to be used in the Gulf Coast region in 1924 and has probably been the most successful of the several geophysical methods employed, though the torsion balance, an extraordinarily sensitive instrument for measuring variations in the value of gravitational attraction, is more accurate for detailed work. The seismic method as applied to salt-plug traps depends on the fact that earthquake waves travel three times as fast through salt as through the adjacent strata. Artificial tremors are induced by exploding charges of dynamite, and the rate at which the resulting elastic waves travel is determined by appropriately located seismographs; those waves that travel for part of their course through a salt mass have the shortest travel times. As a result of growing experience the methods have rapidly improved. Depths can be determined by measuring the reflection of the waves. During the early years of seismic prospecting the effective working depth was 2500 feet, but it has now increased to 15,000 feet or more. The Gulf region has been "reshot" three or four times to locate the favorable oil-bearing structures formed by the deeper salt plugs. Whereas the discovery well at Spindletop was brought in at a depth of 1100 feet, oil is now being obtained from wells as deep as 13,000 feet,[1] and a few exploratory wells, "wildcats" as they are called by oil men, have penetrated to still greater depths, the deepest being nearly 18,000 feet. Geophysical methods were first used in the Gulf Coast region in 1924, and in the next 4 or 5 years as many domes were found as had been found in the preceding 30 years.

The methods of geophysical exploration have been so greatly improved that they can be used to find promising structures of many kinds. They are now being used all over the world in the search for new oil fields. A remarkable development in recent years is the campaign to find oil by these methods under the water of the Gulf of Mexico adjacent to Texas and Louisiana, as far as 50 miles from shore. Several "fields" have so far been found.

The developments in seismic methods for underground exploration illustrate the reciprocal influence of pure and applied science. The seismograph was devised to study the transmission of earthquake waves, without thought of practical application. After it began to be used in a large way in the search for oil in 1924, methods and instruments improved at an accelerated rate. The technique thus developed in response to practical considerations is now being applied to such matters of theoretical interest as the determination of the thickness of glaciers in the Alps and of the ice sheet in Greenland.

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[1] The world's deepest producing well in May, 1948, was in the Wind River Basin, Wyoming, yielding oil from a depth of 14,309 feet.

Oil Reserves of the Nation. Our proved reserves, on January 1, 1948, were 21 billion barrels of oil. By "proved reserves" we mean the recoverable oil underground that can be estimated with considerable accuracy because data are available as the result of drilling. The reserves in an oil field are regarded as proved if (1) the size of the field is known because sufficient drill holes have been put down to outline the field, (2) the thickness of the oil-bearing bed or beds is known, and (3) the volume of the pore space in these beds is known. From these figures the volume of oil in the ground can be computed. However, the recoverable oil, which is the oil that can be brought to the surface by means of present-day practice, is only a fraction of the oil in the ground. The oil that can be recovered, as already stated, amounts to 15 to 40 per cent of the oil in the ground, the amount depending on, among other things, the rate at which the oil is being recovered, a rate that should not exceed a certain optimum.

Since the birth of the oil industry in 1859, 33 billion barrels of oil have been produced. Nevertheless the proved reserves are larger than ever before, the result no doubt of the great incentive to find oil and the remarkable skill that has been developed in finding it. The proved reserves, at the present rate of consumption, 1700 million barrels a year, would last 14 years. However, additions to the reserves are continually being made by new discoveries, though at a slower rate than formerly because the easily found oil has already been found. Moreover, long before the oil fields are exhausted, substitute reserves will be developed. Therefore the oil supply will last much longer than 14 years. As now foreseeable, the substitute reserves will be utilized in the following order: first, natural gas, which will be converted into gasoline; second, coal will be used as the raw material in making gasoline and by-products; and, finally, oil shales. The Green River oil shales of Wyoming and Colorado alone are capable of yielding by distillation 100 billion barrels of oil, three times as much as has already been yielded by the oil wells of the United States from the beginning of the oil industry in 1859.

In contrast to the proved reserves are the undiscovered reserves. The magnitude of these reserves has been estimated, but the figure is of only very general value and is subject to change as new information becomes available. The basis, in brief, is as follows. The United States has an area of nearly one million square miles that is geologically favorable for the occurrence of oil. Based on the performance of the already producing portion of the area, the total yield should be about 100 billion barrels. Having taken out 33 billion barrels, we have still left in the ground about two-thirds of the original amount of oil recoverable by the present means of extraction.

ORE DEPOSITS

In addition to the mineral fuels, modern industry utilizes many other mineral substances. At present the number is nearly one hundred, but it will undoubtedly increase. Research is steadily finding or creating new uses for mineral products, and minerals considered today to be only of scientific interest may tomorrow become of economic value.

Mineral deposits fall naturally into two classes: *metallic* and *nonmetallic*, on account of the very different natures, associations, and modes of origin of the two classes. The metallic or, more accurately, the metalliferous include the deposits of the industrial metals and the precious metals—gold, silver, platinum, and allied metals. The nonmetallic include the mineral fuels, cement materials, ceramic materials, building stones, gems, and a host of others. The nonmetallic substances mined in the United States exceed the metallic threefold in value of annual output, and the trend of the times is toward their greater and greater industrial utilization.

Definition of Ore. An *ore* is a mineral aggregate from which one or more metals can be extracted at a profit. The essence of this definition is in the phrase "at a profit," for "ore" is an economic concept. Consequently many factors influence what is and what is not ore. When, as in 1932, the price of copper sank to 5 cents a pound, far below its long-time average of 13 cents, much copper-bearing material that had been ore fell out of the class of ore. On the other hand, the steadily increasing purchasing power of gold restored to the rank of ore much gold-bearing material that had not been ore during the decade after World War I, and the increase in the price of gold in 1934, from $20.67 to $35 an ounce, added enormous tonnages of gold ore to the world's supply. Advances in metallurgical technique by making it profitable to work material of lower and lower grade also influence the status of what is or is not ore. The development of new uses can transform a worthless material into a valuable ore; for example, the discovery that the metal tungsten is one of the most useful steel-alloying elements has changed the status of the chief tungsten mineral from an opprobrious impurity to a much-sought constituent.

Mineralogically, an ore consists of one or more metalliferous minerals inclosed in a matrix of worthless material. Technically this matrix, which consists of mineral or rock matter, is termed the *gangue* (găng). The proportions of valuable constituent to gangue differ enormously, even in ores of the same metal. Iron ore of the highest grade is made up solidly of the iron-bearing mineral magnetite, but iron ore at the lower limit of commercial availability consists of 30 per cent of magnetite mixed with 70 per cent of gangue. In contrast to the high metal content of iron ores, most gold ores contain 0.001 per cent of gold disseminated through 99.999 per cent of gangue. In general, the gangue of the average ore greatly exceeds in amount the valuable metal it contains. In the preliminary treatment of ores in metallurgical plants the prime purpose is to separate as cleanly as possible the metalliferous constituents from the gangue. By this treatment, a concentrated product is produced which is much higher in metal content than the original ore. The nature of the gangue influences the cost of concentration and subsequent treatment; hence the gangue is generally as important as the metallic content in determining whether a material is ore.

From many mining districts the concentrates are shipped to industrial centers for final treatment: zinc concentrate from Australia to Belgium, tin-bearing concentrate from Bolivia to the United States, literally from the ends of the Earth; and this movement of mineral commodities bulks large in international commerce.

The metallic content of the ores of the precious metals—gold, silver, and platinum—is measured in troy ounces per ton. However, gold ore, because gold by legal enactment has the fixed value of $35 a troy ounce, is commonly cited as carrying so many dollars to the ton. The ores of the industrial metals are generally measured in percentages; for example, an iron ore is said to contain 60 per cent of iron; a copper ore, 1 per cent of copper; and so on.

PRIMARY DEPOSITS

Occurrence and Origin of Ore Deposits. Mineral deposits occurring in bedrock are called *primary deposits*, to distinguish them from those that have been derived from them by erosional destruction, which are called *secondary deposits*. The most important of the secondary deposits are the stream placers, which are stream gravels carrying valuable minerals that have become concentrated in paying amount during transport by the streams.

Primary ore deposits as a rule do not occur singly but are localized in small areas, and these areas determine where the world's mining districts are situated. Many mining districts produce mainly one metal; indeed some districts produce only one metal. Hence mining districts are usually designated by their principal product, as the Rand gold district of South Africa, or the Butte copper district of Montana.

That ore deposits tend to be localized in districts—their gregariousness as we may call it—is of advantage to man in many ways. Among these advantages is the fact that intensive geologic study leads to a knowledge of the idiosyncrasies of the ore bodies of a district. This knowledge is of the highest value in finding extensions of the known ore bodies and in finding the still undiscovered deposits; in fact, this has been the most successful method of ore hunting so far developed.

Most mining districts are situated near, around, or on areas of intrusive igneous rocks. Indeed, this association is so common that, if igneous rocks do not occur in or near a mining district, it is thought by many that they are nevertheless present but concealed in depth; but this is probably an extreme view.

The association of mining districts with igneous rocks early led to the idea that there is a fundamental relation between the igneous rocks and the origin of ore deposits. The fact that at Vesuvius and other volcanoes during times of heightened activity hot gases carrying iron, copper, and lead are given off and visibly form metalliferous minerals, such as galena and hematite, suggested that the magmas from which the igneous rocks were formed supplied the metals in the ore deposits and that the metals were released from the magma at the time it solidified. The magma is the ore bringer.

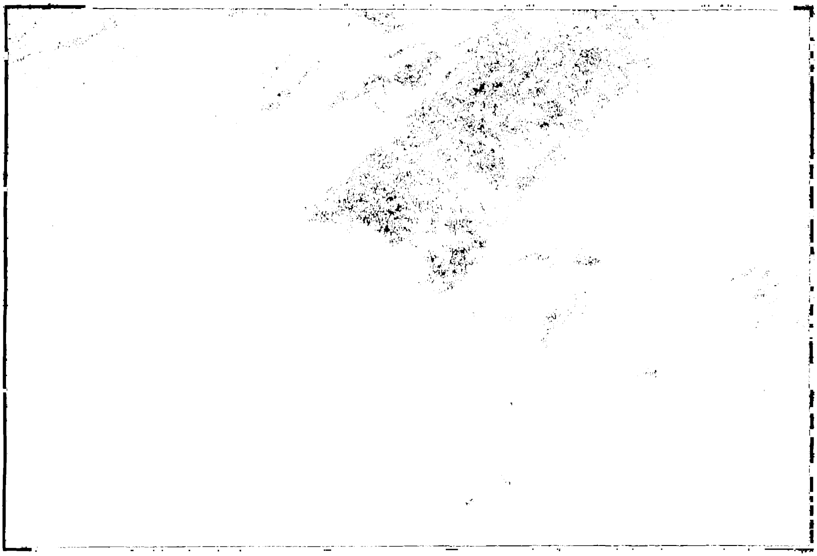
Most ore bodies formed in connection with igneous action are associated with plutonic rocks occurring as stocks and batholiths. Some occur in areas of volcanic rocks, but here the ore-forming solutions as well as the volcanic rocks themselves are both derived from a parent magma that solidified under plutonic conditions at a comparatively shallow depth, possibly a mile or two beneath the surface. In a few mining districts the ore deposits are in volcanic necks. As an empirical fact, ore deposits that are genetically associated with stocks and batholiths have a greater vertical extent than those associated with volcanic rocks. At present the world's record for downward persistence of an ore body formed by solutions derived from deep-seated magma is held by a gold mine in Mysore, India, whose depth is 9200 feet, whereas ore deposits associated with volcanic rocks are likely to give out at depths of 500 to 1000 feet.

The solutions that carry the metals from the solidifying magma are at high temperatures, so high that they are gaseous. As these solutions move up through fissures toward the Earth's surface, they are steadily losing temperature, they are reacting with the wall rocks of the fissures, they are becoming admixed with more or less ground water, and the pressure on them is diminishing. As a result the composition of the solutions changes and certain constituents become insoluble and consequently are deposited in the fissures. Loss of temperature is probably the most potent factor in thus causing precipitation. As the solutions leave the magma they are gaseous, consisting mainly of steam, but during their journey upward they condense to water. Such hot-water solutions are the agents by which most ore bodies have been formed, and the ores thus deposited are said to be of *hydrothermal* origin.

According to this theory of origin every mineral deposit, if explored deeply enough, would be found to give out in depth; in other words, every vein or lode has roots. The theory explains readily the well-known fact that the interiors of large areas of intrusive igneous rocks (batholiths) are barren of mineral deposits. The large size of an area of exposed igneous rock is the result of widespread removal of the rocks that once extended as a roof over the igneous mass. Consequently, as the top of a batholith is roughly dome shaped, the greatest vertical thickness of rock has been removed from over the central portion of the area of igneous rock. In the interior, then, erosion has cut down deeply enough to remove even the roots of whatever veins were once present, toward the borders the roots remain, and at the borders and in the surrounding rocks the veins still have a considerable vertical range.

That veins persist downward in some batholiths, as at Butte, Montana, to depths exceeding 4000 feet, proves that the batholith had solidified to a depth of more than 4000 feet before the shell of rock thus formed began to fracture and allowed the magmatic gases to escape from the still-fluid magma beneath it. The process of fracturing that makes the channel ways for the ascending ore-forming solutions is likely to persist for a considerable time. Recurrent movements may reopen the veins, as already mentioned, and in many districts later fracturing has severed the veins and dislocated the segments. One of the tasks of structural geology is to determine the laws that govern the formation of the channel ways, without which the mineral veins could not have been formed, and to find the faulted segments into which the veins have been severed.

Forms of Ore Bodies. The shape or form of an ore body is of great practical importance, for it determines the way in which the ore body can be worked. A tabular mass of ore more or less sharply delimited from the rock inclosing it is termed a *vein* (Fig. 343). The ideal vein is bounded by well-defined walls; the overhanging wall, as illustrated on p. 375, is the hanging wall, and the other, on which the miner obtains his footing, is appropriately called the foot wall. The ore in the ideal vein breaks cleanly from its walls. It is tacitly part of the definition of a vein that if the vein occurs in other than massive igneous rocks (such as granite) it cuts across the structure of the inclosing rocks. Beliefs in the characteristics of an ideal vein are of great practical significance, for they have greatly influenced the mining laws of the United States; and because the average vein departs widely from the ideal, these departures have caused and are still causing enormous difficulties in squaring the legal view of what a vein should be with what Nature has actually formed.



H. G. Ferguson, U. S. Geological Survey.

FIG. 343. Gold-bearing quartz vein, Alleghany district, California. The sharply defined white quartz vein is well shown, together with the veinlets branching from it.

Some veins are bordered by zones of mineralized rock rich enough to be mined along with the vein filling. These mineralized zones as a rule shade off imperceptibly into rock too poor to mine. Consequently an

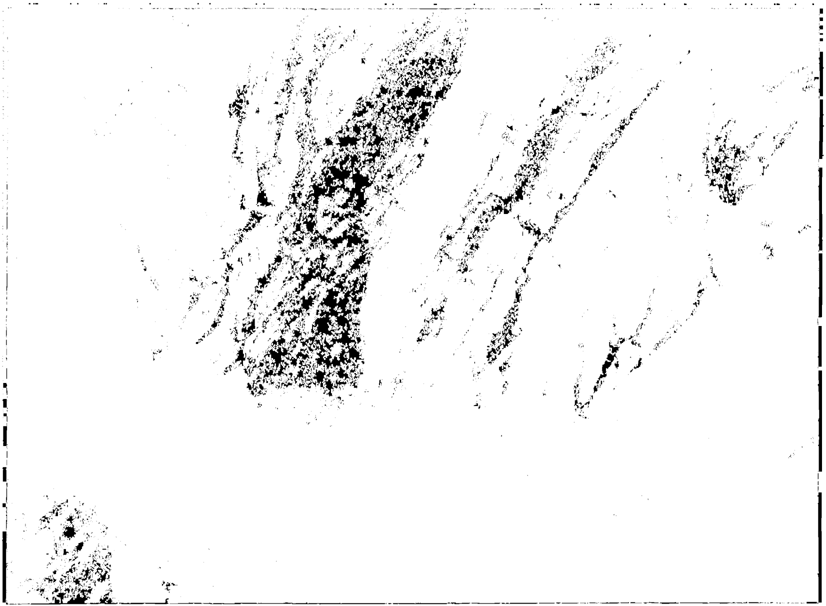
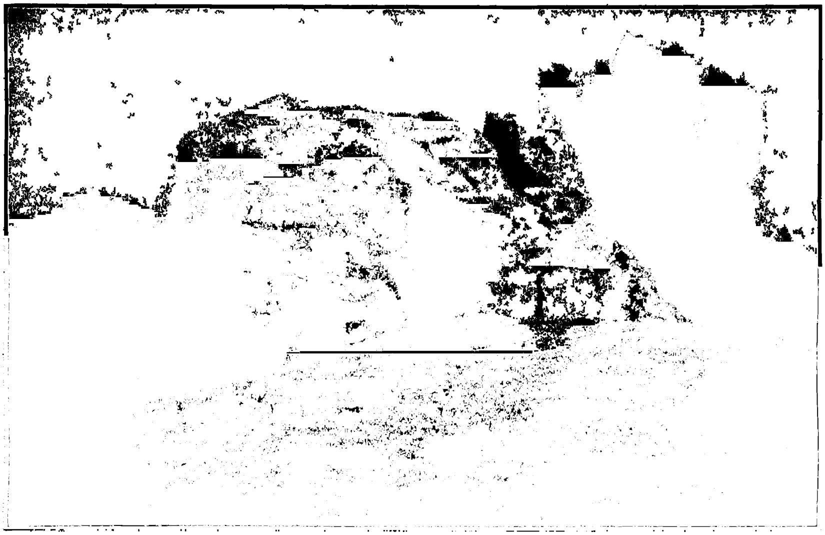


FIG. 344. Gold-bearing lode as seen underground, a member of the Mother Lode system Sierra Nevada gold belt, California.



Adolph Knopf, U. S. Geological Survey.

FIG. 345. Outcrop of a gold-bearing quartz lode. Mother Lode system, Sierra Nevada, California.

ore body of this kind, instead of being bounded by well-defined walls, is bounded by "assay walls," so called because continuous sampling and assaying to determine its metallic content are necessary in finding the limits of the ore body. Therefore such boundaries fluctuate with the price of the metal that is being mined.

A *lode* consists of several veins spaced closely enough so that all of them, together with the intervening rock, can be mined as a unit (Fig. 344). All the larger tabular ore bodies tend to be of this general character, and there is no sharp distinction between a vein and a lode. A spectacularly large outcrop of quartz, part of one of the gold-bearing lodes of the Mother Lode system of California, is shown in Fig. 345.

Some ore bodies in sedimentary rocks conform to the bedding of the inclosing strata. Such ore bodies are either of sedimentary origin or of replacement origin (p. 130). Genetically these two modes of origin are wholly unlike, and the distinction is of great practical importance although not always easy to make. If of sedimentary origin, the ore bodies are beds; if of replacement origin they are termed *bed veins*.

Fissure Veins. Some ore bodies have been formed by the filling of cavities or openings in the crust, others have been formed by replacement, and others have been formed by a combination of both processes. Openings in the crust are made by fracturing and fissuring. As the walls of a fissure are not rigorously plane but are more or less irregular wavy surfaces, movement along them relative to each other causes open spaces to form along the course of the fissure. These openings become the receptacles of vein matter, and the vein thus formed, when followed along its course or down its dip, is found to expand and contract in width—"to swell and pinch," in the language of the miner. The abruptness of this swelling and pinching depends on the original irregularities in the fissure and determines how far the vein departs from the ideal tabular form. By renewal of movement along the vein the fissure may be reopened, and more vein matter deposited in the newly formed openings. In this way many veins have grown in size. The early formed portions of the vein are likely to become crushed during the later movements, and angular fragments of ore that was formed in the early stages may become incorporated in the ore deposited later. The stages of growth of the vein are thus recorded within the vein, and a sequence in the order of deposition of the minerals can often be ascertained. These facts are not only of theoretical interest but also of practical consequence.

The vein-forming solutions that deposit the minerals in the openings along a fissure soak out into the wall rocks. There they bring about many changes, depending on their composition, on their temperature, and on the composition of the wall rocks. The effects may extend from a few inches up to hundreds of feet from the vein. From the study of these effects, conclusions can be drawn as to the composition and temperature of the ore-forming solutions.

Almost universally the ore-forming solutions in soaking outward from the fissures cause crystals of pyrite to grow in the wall rocks. Nearly all ore-forming solutions carry dissolved sulphur, and when these solutions permeate the wall rocks this sulphur combines with the iron of whatever iron-bearing minerals are present to form pyrite, generally as well-crystallized cubes: the wall rock becomes *pyritized*. The pyrite thus formed in the wall rock is a conspicuous, easily recognizable effect of the ore-forming solutions. Such pyrite is of replacement origin, as it has grown in the body of an old mineral as the result of the removal of part of the old mineral and the deposition of sulphur in the space thus made.

Replacement Ore Bodies. More interesting practically is the ability of the ore-forming solutions under some circumstances to transmute the wall rocks into ore. The vein is then bordered on one or both sides by a layer or layers of ore of replacement origin. The resulting ore body is therefore of dual origin: it was formed partly by the filling of open spaces and partly by replacement.

Many ore bodies are wholly of replacement origin. Limestone is the most common rock thus transformed into ore. In some ore bodies of this origin the evidence of replacement is astonishingly impressive. Although the ore consists solidly of sulphides—galena, sphalerite, and pyrite—it preserves faithfully the bedding and other sedimentary features of the vanished limestone. On seeing such an ore body underground we often find it hard to believe that we are viewing an ore body and not merely country rock. In ore bodies of replacement origin, the ores, in contrast to those in fissure veins, were formed by the ore-forming solutions dissolving the country rock and simultaneously depositing in the space thus formed an equal volume of ore. The thoroughness of replacement in many mining districts is impressive, but the immense volume of rock removed during replacement is perhaps even more impressive. The process of replacement is of prime importance to an understanding of ore deposits.

Composition of Ores. In the primary ore bodies of copper, lead, zinc, and silver, the metals were originally deposited in compounds in which they are united with sulphur—in short, as sulphides. Galena (PbS) is the ultimate source of all the world's lead; sphalerite (ZnS) is the chief source of our zinc; chalcopyrite ($CuFeS_2$) and other copper-bearing sulphides are the ultimate sources of our copper. Gold, it is true, is deposited as native metal, but it is almost invariably associated with pyrite (FeS_2), the most widely prevalent of all sulphides.

Weathering of Ore Bodies. Certain important consequences follow from the fact that the primary ore bodies generally contain sulphides. As soon as an ore deposit becomes exposed at the Earth's surface by the progress of erosion, the sulphides in it begin to oxidize—to unite with the oxygen of the atmosphere—and the oxidized products begin to unite with water and carbon dioxide. A host of new minerals is formed. Prominent among these changes are those that overtake the pyrite. Pyrite alters to rusty yellowish brown limonite, which imparts to the outcrops of veins and lodes their characteristic iron-stained appearance. Not only is limonite formed from the pyrite, but also powerful solvents, such as sulphuric acid and ferric sulphate, are generated, and as these solvents trickle downward they attack the other sulphides. They dissolve the silver and copper and cause them to move downward. As a result of these changes the ore body at the outcrop is likely to be vastly different from what it is in depth. It differs in appearance, in composition, and in metallic content.

These changes stop at or near the water table (p. 119), because movement of subsurface water is generally slow below the water table and the supply of oxygen therefore becomes scant. In humid regions the water table is near the surface; in arid regions it may be as much as 2400 feet below the surface. As the position of the water table is influenced also by the relief of the land, the depth to which the oxidized zone extends is a function of climate and topography.

At the water table, then, the character of an ore is likely to change drastically. As a result, the metallurgical processes adapted to the oxidized ore may be wholly unsuited to the primary sulphide ore found below the water table. Many a mine, equipped to treat the ore occurring near the surface, has failed for this reason alone. Moreover, the metallic content of an ore body is likely to become impoverished or enriched, depending on the nature of the chemical reactions that took place in the zone of oxidation. The explanation of why some ores are impoverished and others are enriched involves the application of chemistry to the problems of geology.

Supergene Enrichment. The enrichment that has affected certain ore deposits just below the zone of oxidation is so important that it merits brief mention. Copper illustrates the principle most readily. By weathering, the copper in the outcrop and underlying zone of oxidation is rendered soluble. Consequently it is dissolved by descending water and is carried downward. When the copper-bearing solution comes into contact with the unaltered sulphides that occur in the ore body below the zone of oxidation, it reacts with them and the copper is precipitated as a bluish black film or coating on the sulphides. The copper content of the primary material is thereby increased; and in many places, notably Utah, Nevada, Arizona, and New Mexico, the primary mineralized rock, which was too lean to be ore, has been enriched sufficiently to bring it up to ore grade (Fig. 346). The process is termed *supergene enrichment*. In the States mentioned this process has been effective in forming enormous deposits of copper ore. Although the deposits are low in grade, containing 1 per cent or less of copper, the application of mass-production methods has made them immensely productive and profitable and during the last four decades has revolutionized the copper industry.

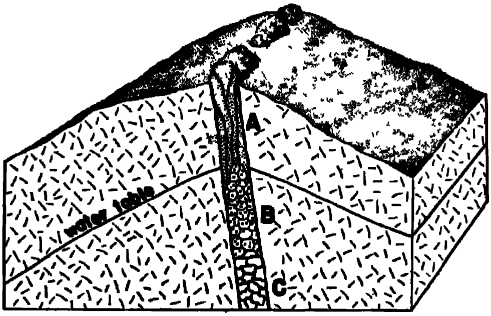


FIG. 346. Quartz vein, showing effects of weathering and supergene enrichment.

A. Oxidized leached ore: limonite and quartz; depleted of copper.
B. Enriched ore: chalcocite, pyrite, and quartz; rich in copper.
C. Primary ore: chalcopyrite, pyrite, and quartz; low in copper.

Iron-Ore Deposits. Most ore deposits, as already outlined in the preceding pages, are concentrations of metals brought about by igneous activity. Other methods of concentration, however, are effective in forming ore deposits of economic value. These other methods can be illustrated most readily by means of the iron-ore deposits.

Iron is by far the most useful of the metals, and fortunately its ores are far more abundant than those of all other metals. This abundance has resulted partly because iron-ore deposits are formed in more ways than those of any other metal, but chiefly because many of them were formed by sedimentation, a process that produces far larger ore deposits than any other process of concentration. If we consider that an iron-ore district containing a reserve of more than one billion (10^9) tons of ore is a district of the first magnitude, then there are ten or twelve such the world over; and in at least seven of them the ore is of sedimentary origin.

It is occasionally said that aluminum in the future will supplant iron as an industrial metal. This assertion is based on the fact that aluminum averages 8 per cent of the Earth's crust, whereas iron makes up only 5 per cent. However, the processes that have concentrated aluminum into ore bodies are few, whereas many have been effective in concentrating iron; moreover, they have operated on a grander scale. As a result, the ores of iron are roughly 40 times as abundant as those of aluminum. Technically and economically it is the ore of the metal that is really important and not the average amount of the metal in the crust. For these and other reasons, chief among which is that "the services of iron, for which there are no substitutes, are well-nigh numberless," iron will remain supreme as an industrial metal.

In the following sketch of the processes by which iron has been concentrated to form commercially valuable ore bodies, four types of deposits are described: first, deposits formed by concentration within molten magmas and therefore formed at high temperature; second, those formed as a result of the iron being carried out of the magma by escaping gases, the iron having been rendered volatile by the presence of chlorine or some other agent, such as those in evidence in the fumaroles at Vesuvius and at the Valley of Ten Thousand Smokes in Alaska; third, the deposits formed by sedimentation; and fourth, those in which the iron has been concentrated by the process of weathering.

Iron-ore bodies formed by concentration within molten magmas are rare. Probably the best-established examples, whose mechanism of concentration can most readily be visualized, are in the Bushveld, Union of South Africa. They are inclosed in gabbro as layers of magnetite that resemble sedimentary strata; they are as much as 15 feet thick and extend for hundreds of miles. At a certain stage while the gabbro magma was solidifying the magnetite began to crystallize, and, being heavier than the magma, the grains sank and accumulated in the form of a layer.

In some districts part of the iron of the magma escaped at the time the magma was solidifying. If the magma was intrusive into limestones, the iron-bearing gases reacted with these limestones. In this way ore bodies of contact-metamorphic origin were formed, as outlined on page 422. If the gases do not meet limestone, they are likely to migrate farther from the magmatic hearth and become condensed to the liquid state. As they move into an environment that is progressively cooler with increasing distance from the magma, they form hydrothermal iron-ore deposits of successively different types. The deposits thus formed carry distinctive minerals that indicate the progressively lower temperatures under which they were formed.

The *minette ores* of Lorraine in France are the world's most valuable ores of sedimentary origin. They consist of oölitic limonite in nearly horizontal beds interstratified with limestone, shale, and sandstone. Although the ore beds are nearly horizontal, they dip westward at a very low angle. They average 30 per cent of iron—a relatively low content—but, because they contain sufficient calcite to form a slag in the furnace, are near supplies of coking coal, and are also near consuming centers, they are of great commercial value. They contain nearly 2 per cent of phosphorus, ordinarily a highly undesirable impurity in an iron ore; but as a result of the development of the Thomas process of smelting in 1879, the phosphorus can be eliminated in the form of a by-product valuable as fertilizer.

By virtue of their sedimentary origin the ore bodies are of large areal extent and of constant composition. Consequently the tonnage of unmined ore, the "reserves," can be estimated with far higher probability than that of ore of any other mode of origin. The reserves are estimated to contain 5000 million tons of ore.

The term minette, meaning "small mine," came to be applied to these ores in this way. At the outcrops the soluble gangue matter (calcite) of the ore as well as the deleterious phosphorus had been dissolved out as a result of weathering. Consequently the ore at the outcrop was enriched in iron and freed from an objectionable impurity. As soon as the water table was reached the primary ore was struck, lower in iron and high in phosphorus. Before 1879 (before the invention of the appropriate process to smelt phosphorus-bearing ores), the primary ores were worthless; hence the outcrop of the Lorraine ores was marked by a line of unimportant small mines—minettes.

During the negotiations at the close of the Franco-Prussian war in 1871, Bismarck was advised by the Director of the Prussian Geological Survey to shift the proposed boundary of Lorraine westward to include the outcrop of the iron ore. This was accepted by the French, who were particularly anxious to obtain concessions of territory near Belfort. At that time neither the Germans nor the French understood the geology and origin of the Lorraine iron ores. The invention of the Thomas process in 1879 made the ores workable and immensely valuable, and the great iron industry of the German Empire was largely based on them. As development proceeded the geologic features of the deposits became clear, and it was seen that although the Germans had the outcrops of the deposits the gentle westward dip caused the iron-ore beds to extend under the soil of France, and that in fact a large fraction of the total reserves lay under French territory. An important iron industry was developed in France on the resources thus unexpectedly left to her. During World War I some of the fiercest, bloodiest drives of that war were undertaken by the Germans to undo the "error" of 1871. By the treaty of Versailles in 1919 *Lorraine annexée* was restored to France, and the entire iron field passed into her possession.

In eastern Cuba are large iron-ore deposits, in which the iron was concentrated to commercial grade by lateritic weathering (p. 48) of serpentine. The ore occurs on a plateau as a blanket 20 feet thick. In the cuts made by mining with steam shovels the ore at the surface can be seen to grade downward into the unaltered serpentine from which it was derived. The serpentine contains 7 per cent of iron; and chemical weathering, by dissolving out almost all the constituents of the serpentine except the iron, has raised the iron content of the residual product to 50 per cent. Such lateritic ores occur also in the Celebes, the Philippines, and elsewhere.

In the Lake Superior region, from which the United States obtains 85 per cent of its supply of iron ore, the ore deposits are the results of a double concentration. First there was laid down by sedimentation an iron formation, consisting of thinly bedded strata of iron-bearing minerals alternating with siliceous strata. In the Mesabi district, which is the most productive iron district in the world, this iron formation averages 25 per cent in content of iron. This is too low in grade to be worked as ore. Fortunately, parts of the iron formation were reconcentrated by weathering, by which the gangue was removed and the iron minerals, because of their insolubility, were left behind. As a result of this purification the residual product contains 50 to 60 per cent of iron and is therefore valuable ore. This weathering took place not under present conditions of climate and topography but under those of remote geologic antiquity, of Pre-Cambrian time.

In conclusion, then, many diverse geologic processes have been effective in concentrating iron into ore deposits of workable grade. Some of these deposits have required a twofold concentration, first by sedimentation and second by weathering.

Gold and Other Metal Deposits. The primary ore deposits of gold and of the industrial metals, copper, tin, lead, and zinc, unlike those of iron, are mainly veins and lodes of hydrothermal origin. Most of these ore deposits are situated either in the border zones of stocks and batholiths of granite or in the rocks that immediately surround such

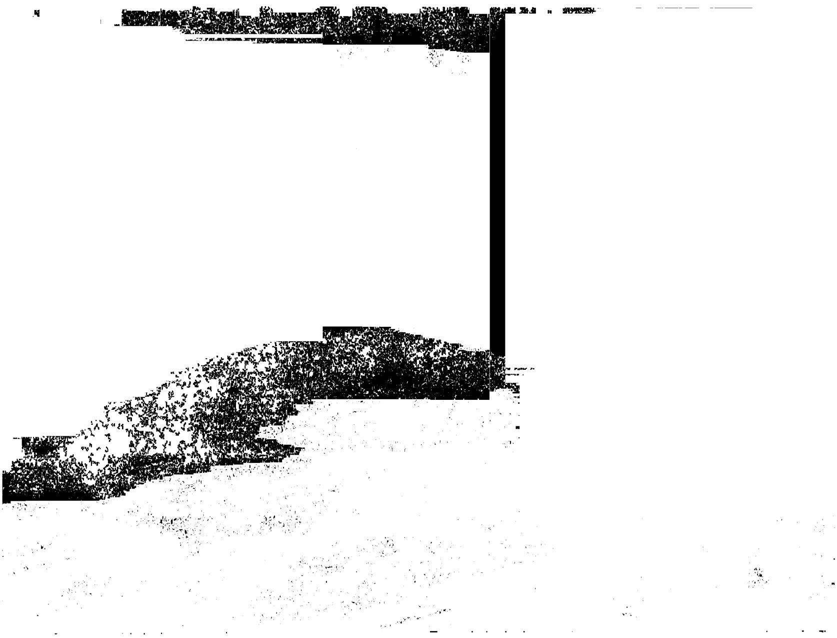
Gold lodes hold the distinction that they are being followed deeper into the crust than are the deposits of any other metal. Depths of more than a mile have been reached in California, 8500 feet in South Africa, and 9200 feet in India. The high temperatures at these great depths and the tendency of the rocks to spall off from the walls of the workings with explosive violence owing to the great pressure are serious handicaps and will probably limit man's quest for metals to a depth of 10,000 feet.

PLACERS

General Features. A *placer* (plăs′er) is a deposit of sand or gravel that contains particles of a valuable mineral in sufficient quantity to be workable at a profit. Gold, tin (in the form of its oxide, cassiterite), platinum, and diamond are the substances commonly won from placers. Most placers are stream gravels. A few, however, are marine beach gravels. Such were the famous gold placers of Nome, Alaska, of which one was the present beach of the Bering Sea, and the others were ancient beaches, formed when the Bering Sea had outlines somewhat different from those it has now. The world's entire supply of the fissionable element thorium is derived from placers; in recent years it has come largely from the beach placers of Travancore in southern India. All heavy, relatively indestructible minerals are likely to become concentrated in placers.

Placers are so easily found and so readily worked, without special equipment or capital, that they have been sought from the beginning of history by the prospector and explorer. Gold placers have been the strongest incentive to this search, and most of our mining districts have been found by the placer miner. The richer placers are soon worked out, and attention then becomes directed to finding the bedrock sources from which the minerals in the placers were derived. If these sources prove to be deposits of economic size, they form the basis for a much more permanent industry; and the ephemeral placer-mining camp becomes transformed into the more staid lode-mining district.

Extraordinary ingenuity has been applied to the problem of working the leaner placers. So brilliant are the results in the more advanced methods of placer-mining—*hydraulicking* and *dredging*—that they are the most efficient forms of mining, in the sense that lower-grade material can be worked by them than by any other method. In hydraulic

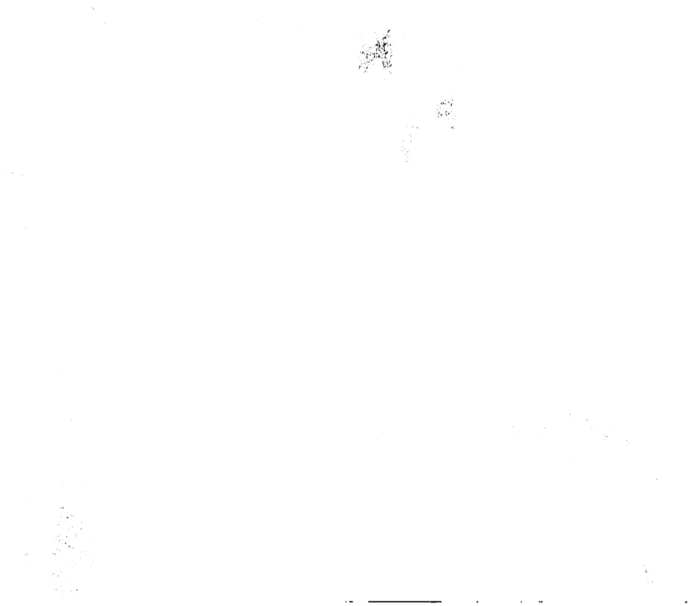


G. K. GILBERT, U. S. GEOLOGICAL SURVEY.

Fig. 347. Hydraulic mining of gold-bearing gravel in the Sierra Nevada. The gravel is being swept by a powerful jet of water into the head of the string of sluice boxes at the extreme left.

mining a jet of water under heavy pressure is directed against a bank of gravel. The gravel thus loosened is swept by a current of water into a long sluice (Fig. 347), where the gold, platinum, or cassiterite, because of its high specific gravity, rapidly settles to the bottom and is caught by appropriate devices at the head of the sluice, and the lighter material, the gravel and sand, is carried through the sluice. In thus concentrating the gold or other valuable material, the miner, using Nature's own method, completes the concentration begun by Nature in forming the placer. In California, where hydraulic mining reached its highest development, gravel containing as little as 3 cents in gold to the cubic yard has been profitably worked.

In dredging, the digging apparatus and the entire concentrating plant are contained on a boat—the dredge (Fig. 348). The dredge floats on a pond; it digs and takes in the gravel at its forward end, and discharges at its stern the tailings—the gravel from which the gold, platinum, or cassiterite has been separated. Thus the pond and the dredge on it gradually travel up the valley. Gold-bearing gravel averaging



BRADFORD WASHBURN.

Fig. 348. Dredge working a gold placer at Fox, near Fairbanks, Alaska.

10 cents in gold to the cubic yard can be worked at a profit by dredging.

At the present time 10 per cent of the world's annual output of gold, 70 per cent of its supply of tin, and much of its platinum are still derived from placers.

Origin of Placers. As soon as the primary ore bodies come into the zone of erosion, they are attacked by weathering. If they contain sulphides, they are subjected to a peculiarly active form of chemical weathering intensified by the sulphuric acid and other powerful solvents formed by the oxidation of pyrite and other sulphides. Only the most resistant substances survive this attack. After being set free from their matrix they eventually find their way by creep and other processes of mass-wasting into the streams. Only the tougher materials withstand the pounding and abrasion to which they are subjected in the stream; the brittle minerals are pounded to minute particles and are swept downstream. The valuable minerals in placers are survivals of the fittest, where fitness is measured by resistance to chemical attack and to abrasion. Hence it is easy to understand why the minerals commonly won from placers are such indestructible substances as gold, platinum, cassiterite, and diamond.

The gold and other valuable minerals occur in particles ranging from the minutest grains up to larger pieces termed *nuggets*. According to the distance that these particles have been carried downstream from their source in bedrock, they are more and more water worn. Angular, jagged, "rough" gold occurs in placers near where it was set free from its matrix in its bedrock source; smooth gold is found farther downstream. It is somewhat surprising, however, to see within how short a distance from their bedrock sources gold particles become smoothly water worn. This rapid wear is due partly to the softness of gold and partly to its great weight, which causes the grains and nuggets to move downstream slowly, with many a halt; and during this slow travel downstream they are pounded and abraded by the passing gravel.

The lowermost few feet of gravel, those resting on bedrock, are much richer in gold than the upper layers of gravel As the miner puts it, the "paystreak" rests on bedrock. Enough gold may occur in cracks in the bedrock to make it profitable to mine several feet of the bedrock below the paystreak. Gold has a truly astonishing ability to sift down into the cracks, joints, and foliation planes of slates and schists that may happen to form the bedrock.

The larger amount of gold on bedrock is due in part to the churning action of streams during high-water stages. The gravel is locally scoured out to great depths during these stages (p. 78), and whatever gold had been distributed through the scoured-out mass of gravel becomes concentrated on the bedrock of the channel thus formed; and as high water subsides, the scour channel becomes filled with barren gravel. Another method of concentration that causes the gold to move toward bedrock is that caused by the tendency of a stream to meander on the gravel-covered floor of its valley. As the stream impinges here and there on the gravel banks, it undercuts them; soon the gravel slumps into the stream and is carried away by the stream, but the gold that was distributed through the vertical bank of gravel from top to bottom has now become concentrated in the stream channel.

It can be seen that a knowledge of the laws governing the transport of detritus by streams and the laws of the origin of land forms is fundamental to understanding how placers are formed. In a region that is in the early stages of the cycle of erosion, the streams are actively at work in downcutting; consequently no gravels are laid down in the beds of the streams more than temporarily. Hence such regions are unfavorable for the occurrence of placers. As a matter of experience it has been found that regions of uplifted and maturely dissected peneplanes are favorable for the development of gold placers. During the peneplane stage the gold is liberated from the veins and veinlets occurring in bedrock throughout the region, and after the region is uplifted and the new cycle of erosion is well under way the gold becomes concentrated in the valleys carved in the old peneplane.

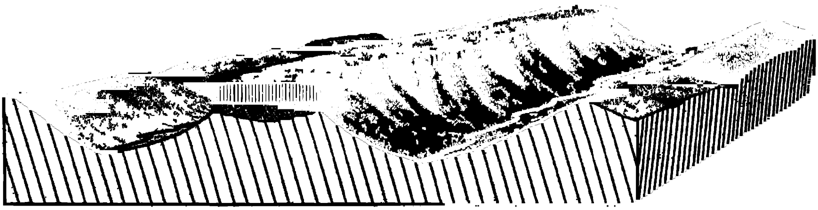


FIG. 349. Ancient stream placer covered by a lava flow; it occupies the divide between two rivers of the present drainage system. Width of block about 3 miles.

Placer deposits have formed at all times during the history of the Earth. Some of the older placers have been buried by later sedimentary rocks; others have been sealed beneath lavas that flowed down the old valleys in which the gravels were accumulating. In regions in which placers have been covered by flows of lava, the streams that formed the placers become dispossessed of their channels and seek new ones. In the course of time a complete revolution in the topography is effected, and the placers capped by the resistant lava flows become the divides between the existing streams. What were once the valley bottoms are now the mountain tops (Fig. 349). Such dead rivers, whose gravels are sealed under lava flows, occur high above the present streams in various parts of the world, notably in the Sierra Nevada, where they have yielded much gold.

Although the finding of placers has generally been the prelude to the discovery of veins and lodes in bedrock, it by no means follows that rich placers are necessarily derived from rich deposits in bedrock. The contrary is almost true. For example, the larger part of the world's platinum has been won from placers in the Ural Mountains, but no workable primary deposits have been found. The platinum in its bedrock home in the Ural Mountains is disseminated through large bodies of rock, which are far too lean to be worked at a profit. The processes by which placers are formed are so efficient, however, that the small amount of platinum distributed through immense volumes of rock has become concentrated in the gravels of the streams that drain the Ural Mountains. Cubic miles of rock were ground up by erosion, and the platinum in them has accumulated in workable amount in the stream beds. Similarly, rich gold-bearing gravels have accumulated in areas in which the gold occurs in bedrock in the form of small auriferous quartz veinlets. The same disparity between the richness of the placers and the poverty of the bedrock sources is even more generally true of tin deposits and is reflected in the fact that 70 per cent of the annual supply of tin is still won from placers.

One of the richest gold-placer regions ever found was on the lower western slope of the Sierra Nevada. Its discovery in 1848 caused the greatest human migration in history. The gravels of the present streams have yielded $900,000,000. Their great richness was the result of a triple concentration of the gold. Near the end of Jurassic time the crust beneath the site of the present Sierra Nevada was invaded by vast volumes of granitic magma. During the solidification of the molten magma, countless small gold-bearing veinlets were formed in the surrounding rocks. This was the first concentration of the gold. A long period of erosion then set in, during which the region was reduced to a land of moderate relief. In the streams that flowed on this old land surface great masses of gravel accumulated, hundreds of feet thick in the larger valleys, and rich in gold in the layers resting on bedrock. The gold thus accumulated was the result of the second concentration. Great eruptions burst out and overwhelmed the streams with volcanic debris and sealed them under lava flows. The streams were forced out of their courses and began to dig new ones. Faulting along the eastern flank of the Sierra lifted the range high into the zone of erosion and tilted the western slope to the west, thereby increasing enormously the erosive powers of the westward-flowing streams. The great canyons that gash the western slope of the range were formed. Where the streams cut through the gravels of the old river beds, they robbed them of their gold, and to this store of easily won metal they added the gold that erosion was liberating from the veins and veinlets in bedrock. As the result of these three stages of concentration the present streams of the Sierra Nevada acquired the extraordinary richness in gold that made them famous the world over.

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APPENDIX A

MINERALS

INTRODUCTION AND DEFINITION

Minerals compose the rocks of which the Earth's crust is built. They are also the constituents of ore bodies and other deposits of economic value. Rocks are the primary documents of geology, and in order to read these documents it is necessary to be able to recognize their constituents. A mineral can be defined as a naturally occurring substance that has a distinctive set of physical properties and a composition expressible by a chemical formula.

Minerals are composed of chemical elements. A few consist of single elements, such as native gold and silver, as these metals are termed when they occur in elementary form in Nature, or diamond and graphite, both of which are crystalline forms of the element carbon. Diamond and graphite illustrate in the most striking way possible what is meant by a mineral. The two are of identical chemical composition, yet each has its own distinctive physical properties: diamond is transparent and is the hardest substance known, whereas graphite is opaque and is among the very softest of substances. Most minerals, however, are made up of two or more elements united in such a way that the product differs in its properties from any of the elements composing it.

There are 96 chemical elements. Less than half of them are abundant; in fact, more than 99 per cent of the Earth's crust is composed of 14 elements. These elements, with their chemical symbols, are shown in the following table in the order of their abundance in the crust:

	Per Cent		Per Cent
Oxygen (O)	46.46	Titanium (Ti)	.62
Silicon (Si)	27.61	Hydrogen (H)	.14
Aluminum (Al)	8.07	Phosphorus (P)	.12
Iron (Fe)	5.06	Carbon (C)	.09
Calcium (Ca)	3.64	Sulphur (S)	.06
Sodium (Na)	2.75	Chlorine (Cl)	.05
Potassium (K)	2.58	All others	.68
Magnesium (Mg)	2.07		
		Total	100.00

Three elements not included in the preceding list are added here because they occur in certain important ore-forming minerals. Each of these elements makes but a minute fraction of 1 per cent of the Earth's crust.

Copper (Cu)
Lead (Pb)
Zinc (Zn)

The minerals that have formed by various combinations among these seventeen elements are very numerous, but those that are abundant, as components either of rocks or of ores, are relatively few, about twenty-five.

PROPERTIES OF MINERALS

CHEMICAL COMPOSITION

A few minerals have a fixed chemical composition; but most of them have a variable composition which, however, can be expressed by a chemical formula. Quartz, one of the most abundant of minerals, has a fixed composition, written SiO_2, which is a sort of chemical shorthand saying that one atom of silicon is united with two atoms of oxygen. Sphalerite, from which most of the world's zinc is obtained, is a mineral of variable composition, which is indicated by writing its formula thus: $(Zn,Fe)S$—thereby indicating that in this mineral atoms of iron proxy for atoms of zinc.

The various minerals react differently to chemical reagents, and these reactions are one means of identifying minerals. It is beyond the scope of this discussion to explain how minerals are identified by their chemical behavior, but many textbooks of mineralogy treat the subject fully.

PHYSICAL PROPERTIES

Crystals. Nearly all minerals are *crystalline;* that is to say, the atoms of which they are built are organized in definite geometric arrangements. A few minerals are *amorphous* (noncrystalline). Under favorable conditions most minerals form *crystals.* Crystals are solids that are bounded by smooth plane surfaces called *faces,* whose arrangement is related to the internal structure of the mineral. As a rule, the crystals of any particular mineral have similar forms of crystallization. For instance, the mineral pyrite crystallizes characteristically in cubes (Fig. 350). Garnet is common as twelve-sided crystals called dodecahedrons (Fig. 351). These crystal forms are characteristic of these minerals, and recognition of the crystal forms aids greatly in identifying the minerals.

Structure. The *structure* of minerals generally refers to their outward shape and form. The following descriptive terms are used in this connection, some of which are self-explanatory: *crystallized,* occurring as crystals or showing crystal faces; *massive,* not bounded by crystal faces—the antithesis of crystallized; *columnar; fibrous* (Fig. 352); *botryoidal* (Fig. 353), consisting of small rounded forms like closely bunched grapes; *reniform,* kidney-shaped; *micaceous,* occurring in thin cleavable sheets or flakes; *granular,* in aggregates of coarse to fine grains; *compact; earthy; oölitic,* formed of small spheres which resemble fish roe.

FIG. 350. Cubic crystals of pyrite. *Striae* on faces are well shown; crystals are intergrown.

Cleavage and Fracture. The manner in which many minerals break is so characteristic that it is of great help in identifying them. If they break so that smooth plane surfaces are produced, they are said to have a *cleavage.* This cleavage invariably occurs along planes, but these planes may or may not be parallel to crystal faces. Some minerals have but one cleavage, others two, three, or even six different cleavage directions. The number of cleavage directions that a mineral has serves as an aid in determining the mineral. Good examples are the cubic cleavage of galena, whereby the mineral cleaves in three planes at right angles to one another, so that it breaks up into innumerable small cubes (Fig. 354); the rhombohedral cleavage of calcite—three planes not at right angles, so that the resulting cleavage fragments are rhombohedrons (Fig. 355); and the cleavage of mica—in one direction only, the most remarkable example of cleavage in the whole mineral

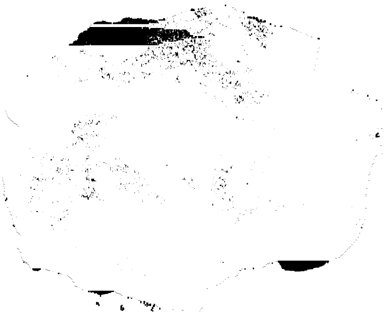

FIG. 351. Dodecahedral crystals of garnet imbedded in mica schist.

kingdom, by virtue of which the mica can be cleaved into sheets or flakes of indefinite thinness.

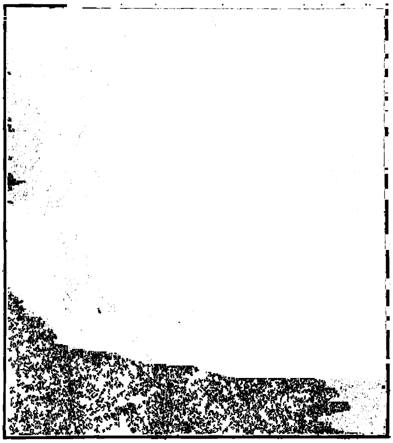

FIG. 352. Fibrous structure, as shown by asbestos.

If a mineral has no cleavage, the nature of its broken surface—its *fracture*—is more or less distinctive. The kinds of fracture are: *conchoidal* (Fig. 362), if the surface of fracture is curved like the interior of a clam shell; *fibrous* or *splintery*, if it is like that of wood; *uneven* or *irregular*, if the surface is rough.

Color. The color of a mineral is one of its most conspicuous physical properties. The color of some minerals is a definite and constant property and serves as a ready means of identification. For example, the golden yellow of chalcopyrite, the lead-gray of galena, the black of magnetite are striking properties of these minerals. Surface alterations are likely to change the color of a mineral, as is shown by the golden tarnish frequently seen on pyrite. To observe the true color of a mineral a fresh surface should therefore be examined.

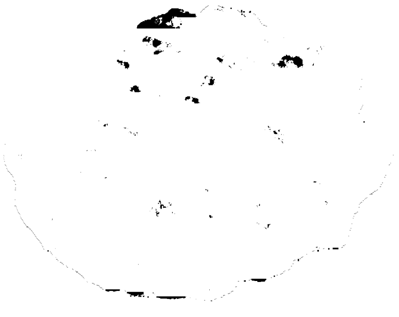

FIG. 353. Botryoidal structure, as shown by pitchblende, the chief ore mineral of radium and uranium.

Many minerals differ in color in different specimens. This is due chiefly to a change in composition, such as an increase of iron content in sphalerite, with consequent darkening of the color of the mineral; or to impurities, such as the red color given to quartz by an admixture of hematite. Other minerals, such as fluorite, show a wide range in color without perceptible changes in composition.

Color of Powder or Streak. *The color of the streak* is an important aid in identifying some minerals. The streak is a thin layer of the powder of the mineral obtained by rubbing the mineral on an unglazed porcelain plate known as a streak plate. The color of the streak may be similar to the color of the mineral or quite different. For example, some varieties of hematite are brilliantly black, but they give a red-brown streak which positively identifies them as hematite.

Luster. The *luster* of a mineral is due to the quality and intensity of the light it reflects. Luster must not be confused with color, for two minerals of the same color can have totally different lusters, just as a black paint of shiny finish, such as an enamel, differs in appearance from a black paint of dull finish because it reflects light differently.

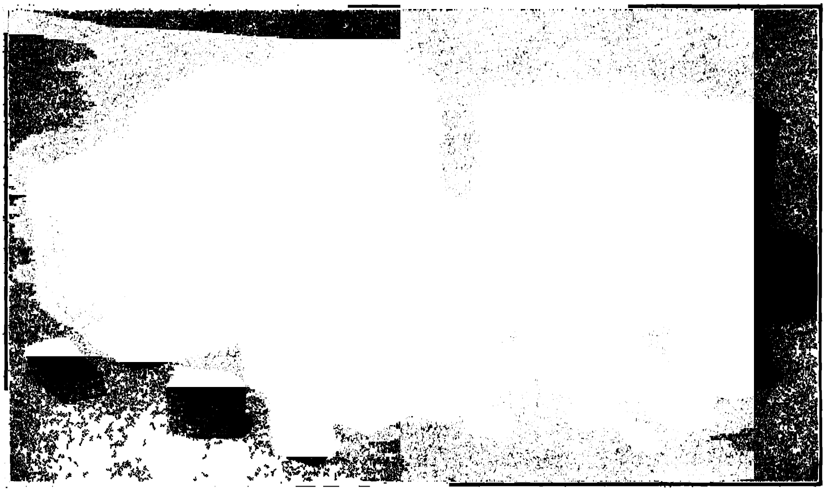

FIG. 354. Cleavage fragments of galena, showing perfect cubic cleavage. Smaller fragments, at right, are reflected in the perfect cleavage face of the larger fragment behind them.

Descriptive terms are given to the various lusters. A list that includes the more important is as follows:

Metallic. Having the luster of a metal. Example: pyrite. Most minerals that give a dark or black streak have metallic luster.

Vitreous. Having the luster of glass. Example: quartz.

Resinous. Having the luster of yellow resin. Example: sphalerite.

Pearly. Having the iridescence of pearl. Example: some varieties of feldspar.

Greasy. Appearing as if covered with a thin layer of oil. Example: some varieties of massive quartz.

Silky. Like silk, as the result of a finely fibrous structure. Example: fibrous gypsum.

Adamantine. Having a brilliant, almost metallic, luster like that of a diamond.

Hardness. Minerals differ greatly in their hardness; consequently the determination of this property is an important aid in identifying them. The relative hardness of a mineral can be determined by comparing it with a series of minerals that has been chosen as a standard scale. The scale consists of the following minerals, each mineral being harder than those that precede it.

SCALE OF HARDNESS

1. Talc
2. Gypsum
3. Calcite
4. Fluorite
5. Apatite
6. Orthoclase
7. Quartz
8. Topaz
9. Corundum
10. Diamond

The relative hardness of a mineral in terms of this scale is determined by finding which of these minerals it can scratch and which it can not scratch. In determining hardness the following precautions must be observed. A mineral softer than another may leave a mark on the harder one which can be mistaken for a scratch. The mark can be rubbed off, however, whereas a true scratch is permanent. Some minerals are commonly altered on the surface to material much softer than the original material. The physical structure of a mineral may prevent a correct determination of its hardness. For instance, if a mineral is powdery, finely granular, or splintery in its structure, it can apparently be scratched by a mineral much softer than itself. When making the hardness test, it is always advisable to confirm the test by reversing the procedure, that is, by rubbing the unknown on the material of known hardness.

FIG. 355. Cleavage rhomb of calcite.

The following materials serve as additions to the above scale. The convenient finger nail is a little over 2 in hardness, as it can scratch gypsum but not calcite. A copper coin is slightly more than 3 in hardness, as it can scratch calcite but not fluorite. The steel of an ordinary pocketknife is just over 5, and ordinary glass has a hardness of 5.5.

Specific Gravity. The specific gravity of a substance is expressed as a number that indicates how many times heavier a given volume of the substance is than an equal volume of water. Minerals range in specific gravity between 1.5 and 20.0; most range between 2.0 and 4.0. There are various instruments that enable one to determine the specific gravity of a mineral with accuracy, but ordinarily it is sufficient to judge the weight of a fair-sized piece in the hand. After some experience rather small differences in specific gravity can be detected in this way, and the specific gravity of a mineral can be roughly estimated.

COMMON MINERALS

The more widely occurring minerals are described on the following pages. The student should compare these descriptions with as many different specimens of the minerals as possible, and should note the form, color, and luster of each sample and make the simple tests for hardness, streak, and specific gravity.

Magnetite. An oxide of iron, a combination of ferrous and ferric oxides, $FeO \cdot Fe_2O_3$, or Fe_3O_4.

Physical Properties. Black; metallic luster. Streak black. Hardness 6. Sp. gr. 5.17. Strongly magnetic; hence its name. Granular or massive; fairly common in octahedral crystals (Fig. 356).

FIG. 356. Octahedron, as viewed from above.

Occurrence. A valuable iron-ore mineral, containing 72 per cent of iron. It is mined in the Adirondacks, New Jersey, Pennsylvania, Utah, and in many other parts of the world. It is common as a minor constituent in rocks, particularly in the darker-colored igneous rocks. The black sand of the seashore is largely magnetite.

Hematite. The ferric oxide of iron, Fe_2O_3.

Physical Properties. Dark steel-gray to iron black; brilliant metallic luster (except in earthy specimens). Streak light to dark red-brown (not clearly apparent in some micaceous specimens). Streak distinguishes it from limonite. Hardness 5.5 to 6.5, but not always determinable. Sp. gr. about 5. Granular, micaceous, earthy (in this form it is red), botryoidal, or reniform. Rarely in crystals.

Occurrence. Hematite is widely distributed in rocks and is the most abundant ore mineral of iron. More than nine-tenths of the iron produced in the United States comes from ores consisting of this mineral. The chief iron-producing districts are near the shores of Lake Superior in Michigan, Wisconsin, and Minnesota. Other important districts are in northern Alabama and eastern Tennessee. Earthy hematite is the pigment that gives many sandstones their red color. It is used in red paints and as a polishing mat

Limonite.[1] Mainly minutely crystalline goethite, $HFeO_2$.

Physical Properties. Dark brown to nearly black. Streak yellowish brown, which distinguishes it from hematite, though both minerals may be present in the same specimen. Hardness 5 to 5.5 (but earthy specimens appear to be much softer). Sp. gr. about 4. Commonly as botryoidal and related forms having radiating fibrous structure; also in stalactitic forms resembling icicles; earthy. Some limonite is a product of chemical alteration of hematite, and structure can not be used to distinguish one from the other.

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[1] Commonly used as a field term for any hydrous iron oxide. See page 32.

Occurrence. A valuable source of iron. Limonite is a common mineral formed by the oxidation and hydration of the iron in previously existing minerals. Ordinary iron rust is limonite. It gives brown, orange, and yellow colors to many weathered rocks, nonweathered sedimentary strata, and soils.

Pyrite. Iron sulphide, FeS_2.

Physical Properties. Pale brass-yellow, but some specimens are tarnished to deeper shades of yellow. Streak black. Hardness 6 to 6.5 (unusually hard for a sulphide). Sp. gr. about 5. Generally granular. Common as crystals, especially as cubes whose faces are marked with fine parallel lines, or *striae* (Fig. 350).

Occurrence. The most common sulphide mineral. Occurs in many rocks and is an important vein mineral. May carry small amounts of gold or copper and so become an ore for both these metals. Is not used as an ore of iron, but as a source of sulphur in the manufacture of sulphuric acid. Its presence in building stones detracts from their value, as its oxidation produces not only iron oxide stains but also sulphuric acid, which causes the stones to disintegrate.

Chalcopyrite (Copper Pyrite). Copper-iron sulphide, $CuFeS_2$.

Physical Properties. Golden yellow; generally tarnished to bronze or iridescent colors. Streak greenish black. Hardness 3.5; hence much softer than pyrite. Sp. gr. 4.1 to 4.3. As a rule massive, rarely in crystals.

Occurrence. An abundant and valuable ore mineral of copper. Occurs widely distributed in vein deposits with many other sulphide minerals.

Sphalerite. Zinc sulphide, $(Zn,Fe)S$.

Physical Properties. Commonly yellow-brown to dark brown, being darker in the varieties containing more iron. Resinous to submetallic luster. Hardness 3.5 to 4. Sp. gr. about 4. White to yellow and brown streak, of lighter shade than the mineral itself. Has brilliantly flashing cleavage planes trending in six different directions. As a rule massive.

Occurrence. The most important source of zinc. Widely distributed, generally in veins or irregular bodies in limestone. Associated generally with galena, pyrite, and chalcopyrite.

Galena (gȧ-lē′nȧ). Lead sulphide, PbS.

Physical Properties. Lead-gray. Bright metallic luster. Streak grayish black. Hardness 2.5 (soft). Sp. gr. about 7.5. Perfect cleavage in three planes at right angles to each other, forming cubes (not apparent in finely granular specimens). Occurs in natural cubic crystals (Fig. 354), but massive and granular aggregates are more common.

Occurrence. The chief source of lead. Some contains silver and serves as an ore of that metal. Commonly occurs with zinc minerals.

Calcite. Calcium carbonate, $CaCO_3$.

Physical Properties. Generally white or colorless. Also variously tinted, gray, red, green, and blue. Usually opaque or translucent; rarely transparent. Hardness 3. Sp. gr. 2.7. Perfect cleavage in three planes at oblique angles to each other, giving rhombic-shaped faces (rhombohedral cleavage; Fig. 355). In crystals generally of rhombohedral form. Effervesces freely on application of a drop of cold acid. This serves to distinguish calcite from dolomite, $CaMg(CO_3)_2$, another common carbonate, which does not effervesce under these conditions.

Occurrence. A very common mineral. Is the chief constituent of limestones and marbles; also common in veins. Used in the manufacture of lime, plaster, and cement, as a metallurgic flux, and in chemical industries.

Dolomite. Carbonate of calcium and magnesium, $CaMg(CO_3)_2$.

Physical Properties. Generally white or gray; rarely flesh-colored. Opaque to translucent. Hardness 3.5 to 4 (harder than calcite). Perfect cleavage in three planes not at right angles to each other (rhombohedral cleavage). Sp. gr. about 2.8. Vitreous to pearly luster. Does not effervesce when a drop of cold acid is applied, unless the acid is placed on a scratched or powdered surface. In this respect it differs from calcite. In granular masses and in crystals, some of which have curved faces.

Occurrence. Composes the rocks called dolomite and dolomitic marble. Also as a vein mineral. As rock, is used as a building and ornamental stone; for the manufacture of some cements; as a source of the metal magnesium and of magnesia for refractory substances; and as agricultural lime.

Gypsum. Hydrous calcium sulphate, $CaSO_4 \cdot 2H_2O$.

Physical Properties. Usually white or colorless. Hardness 2 (easily scratched with the finger nail). Sp. gr. about 2.3. Has one very perfect cleavage; another imperfect cleavage visible in some specimens. In tabular diamond-shaped crystals or in granular masses; also fibrous.

Occurrence. Is a mineral widely distributed in sedimentary rocks, frequently in thick beds. It is commonly interstratified with limestones and shales. Generally occurs in association with salt beds. Is chiefly used for the production of plaster of Paris.

Halite (Common Salt). Sodium chloride, NaCl.

Physical Properties. White or colorless. Hardness 2.5. Sp. gr. about 2.1. Perfect cleavage in three planes at right angles to one another (cubic cleavage). Transparent to translucent. Salty taste. Generally in cubic crystals or in masses showing cubic cleavage.

Occurrence. In thick beds interstratified with sedimentary rocks and associated with gypsum. Used for cooking and as a preservative; also extensively in the chemical industry.

Quartz. Silicon dioxide, SiO_2.

Physical Properties. Colorless or white; but many varieties are colored by impurities: yellow, red, pink, amethyst, green, blue, brown, black. Vitreous luster. Transparent to opaque. Hardness 7. Sp. gr. 2.65. Conchoidal fracture. Commonly in hexagonal crystals similar to Fig. 357. The triangular faces at the ends of the crystals are usually smooth, whereas the rectangular faces between the ends are horizontally striated. Generally massive, as in quartz veins.

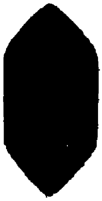


FIG. 357. Model of a quartz crystal.

Varieties. There are many varieties of quartz to which different names are given. A few are: *rock crystal,* colorless quartz, commonly in distinct crystals; *amethyst,* quartz colored purple or violet; *rose quartz,* usually massive with a pink color; *smoky quartz,* smoky yellow to brown or almost black; *chalcedony,* finely fibrous variety, translucent with waxy luster; *agate,* a variegated chalcedony delicately banded in different colors.

Occurrence. Quartz is one of the most common minerals. It is abundant in many rocks and is the commonest vein mineral. It makes up the largest part of most sands. It is widely used in its various colored forms as ornamental material. It is used as an abrasive, in the manufacture of glass and porcelain, in paints, scouring soaps. As sand it is used in mortars and cements. Quartzite and sandstone—rocks made up largely of quartz—are used in building.

Garnet. Garnet is a group name, there being several kinds, which differ from one another in the elements they contain. All are silicates of analogous formulas. The most widely prevalent garnet, almandite, contains ferrous iron and aluminum, $Fe_3Al_2(SiO_4)_3$. Other garnets contain magnesium, calcium, manganese, and ferric iron.

Physical Properties. Color varies with the composition. Most commonly red or brown. Also yellow, white, green, black. Transparent to almost opaque. Hardness 7. Sp. gr. 3.2–4.3 (varies with composition). Generally well crystallized, either in a form showing 12 rhombic-shaped faces (dodecahedron, Fig. 351) or 24 trapezium-shaped faces (trapezohedron).

Occurrence. Garnet is a widely distributed mineral, occurring most commonly in metamorphic rocks. Used as an inexpensive gem stone and, because of its hardness, as an abrasive material.

Orthoclase (Potassium Feldspar). Potassium-aluminum silicate, $KAlSi_3O_8$.

Physical Properties. Colorless, white, gray, flesh-colored, pink, and red; rarely blue-green and green. Hardness 6. Sp. gr. 2.56. Has two good cleavages making angles of 90° with each other (whence the name of the mineral).

Occurrence. The most common silicate. Widely distributed as a prominent constituent of rocks of many kinds, but most abundantly in granite and allied rocks. Also in large crystals and cleavage masses in coarse-grained rock known as pegmatite. From these pegmatites it is quarried in large amounts for use in the manufacture of porcelain.

Plagioclase (plă'jĭ-ō-klāz) **Feldspars.** Sodium-calcium-aluminum silicates.

Physical Properties. Various shades of gray, less commonly white. Transparent to opaque. Hardness 6. Sp. gr. 2.6–2.76. Have two cleavages making angles of nearly 90° with each other, one of them (the basal cleavage) being better than the other. Most plagioclase can be distinguished from orthoclase by the presence on the basal cleavage planes of a series of striae (fine parallel lines, which resemble rulings made by a fine diamond point). Some cleavage surfaces, especially of dark-gray plagioclase, give a beautiful play of colors when specimen is rotated in good light. The white variety commonly occurs in thin-bladed crystals with curved surfaces and a pearly luster.

Occurrence. In much the same manner as orthoclase. The plagioclase in gabbros is likely to be dark colored or black and therefore not easily distinguishable from the associated pyroxene.

Muscovite (White Mica). A complex silicate containing potassium and aluminum.

Physical Properties. Has a perfect cleavage in one direction, which allows the mineral to be split into exceedingly thin sheets or flakes. These sheets are flexible and elastic. Transparent and almost colorless in thin sheets, but impurities commonly form dark splotches. In thicker blocks is opaque and is colored in light shades of brown and green. Hardness 2 to 2.5. Sp. gr. 2.76–3.

Occurrence. A common rock-making mineral. It occurs in granite together with quartz and feldspar, and with the same minerals in pegmatite. It is characteristic of a series of rocks made up of abundant mica, in which it is arranged in parallel orientation, with the result that the rocks split in flakes and slabs parallel to the cleavage of the mica. These rocks are known as *mica schists.* Is used chiefly as an insulating material in the manufacture of electrical apparatus. There are many minor uses.

Biotite (Black Mica). A complex silicate containing potassium, magnesium, iron, and aluminum.

Physical Properties. Perfect micaceous cleavage. Cleavage sheets and flakes are flexible and elastic. Generally dark green, brown, or black. Thin sheets usually have a smoky color (differing thus from the almost colorless muscovite). Hardness 2.5 to 3. Sp. gr. 3.0.

Occurrence. An abundant rock-making mineral, common in granites and many gneisses and schists.

Chlorite. A complex silicate containing magnesium and aluminum. A numerous group of minerals of similar properties are called collectively *chlorite* because of their prevailing green color.

Physical Properties. Perfect micaceous cleavage. Flakes are flexible, but are not elastic (differing in this lack of elasticity from muscovite and biotite). Green of various shades. Hardness 2 to 2.5. Sp. gr. 2.65–2.96.

Occurrence. A common rock-making mineral. The green color of many rocks is due to the presence of this mineral. This is particularly true of many schists and slates (green roofing slates).

Serpentine. Hydrous magnesium silicate, $Mg_3Si_2O_5(OH)_4$.

Physical Properties. Olive green, yellow-green to blackish green. Luster greasy or wax-like; silky when fibrous. Hardness 2.5 to 5, generally 4. Sp. gr. 2.50–2.65. Usually massive but also fibrous or felted.

Occurrence. A common mineral, widely distributed. Invariably an alteration product of some magnesian silicate, chiefly olivine. It is the chief constituent of the rock called serpentine, some varieties of which are used as ornamental stone. The fibrous variety known as *chrysotile* is the principal source of asbestos.

Olivine. Silicate of magnesium and iron, $(Mg,Fe)_2SiO_4$.

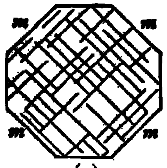
Physical Properties. Olive green to yellowish green; rarely brownish. Transparent to opaque. Hardness 6.5 to 7. Sp. gr. 3.27 to 3.37

Vitreous luster. Conchoidal fracture, causing it to look as if it were a yellowish green quartz.

Occurrence. In dark igneous rocks—gabbros, peridotites, and basalts.

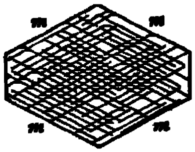
PYROXENE AND AMPHIBOLE

These two abundant rock-making minerals are similar in some respects, and consequently it is difficult to discriminate them in most rocks, where good crystal forms are rare. However, it is well to study them separately under favorable conditions, in order to appreciate their differences as well as their points of similarity.



(*a*)

FIG. 358*a*. Cross-section of pyroxene (eight-sided) perpendicular to the long axis of the crystal. The mineral cleaves along planes parallel to the faces indicated by *m*. Alternate faces (those marked *m* and those unmarked) are practically at right angles to each other.



(*b*)

FIG. 358*b*. Cross-section of amphibole (six-sided) perpendicular to the long axis of the crystal. The mineral cleaves along planes parallel to the prism faces (*m*), which make angles of 56° and 124° with each other.

Pyroxene. Silicate of calcium and magnesium; also varying amounts of aluminum, iron, and sodium. Pyroxene is the name of a group of minerals comprising many varieties dependent on chemical composition.

Physical Properties. Light to dark green or black; rarely white. Commonly opaque. Hardness 5 to 6. Sp. gr. 3.1 to 3.6 (varies with composition). In prismatic crystals with eight sides (Fig. 358*a*); in reality square prisms whose corners are truncated. The angle between alternate faces is therefore nearly 90°. These faces will fit into the corner of a box or tray. By means of these angles pyroxene can best be told from amphibole. Some specimens show a fair cleavage parallel with the faces lettered *m* in the figures, the angle between the cleavage faces being also nearly 90°. This cleavage is not visible in all specimens used for demonstration purposes.

Occurrence. Pyroxene is a highly abundant rock-making mineral occurring chiefly in dark-colored igneous rocks. In these it is dark colored or black; the light-colored or white pyroxene is restricted to the metamorphic rocks. Rare in rocks that contain much quartz.

Amphibole. Silicate of calcium and magnesium with varying amounts of aluminum, iron, and sodium. Somewhat like pyroxene in composition, but contains water of constitution, *Physical Properties.* Light to dark green or black; rarely white. Commonly opaque. Hardness 5 to 6. Sp. gr. 2.93 to 3.8 (varies with composition). Commonly in prismatic crystals with six sides. Figure 358*b* shows that the angles between the faces lettered *m* are 124° and 56° (very different from the corresponding angles in pyroxene). Has a good cleavage parallel with the faces lettered *m*.

The differing forms of the crystals, the differing cleavage angles, and the fact that amphibole has the better cleavage are the chief distinctions between amphibole and pyroxene. If cross-sections can be found, amphibole, being six-sided, can readily be distinguished from pyroxene, which is eight-sided. Compare Fig. 358*a* with 358*b*. Amphibole as a rule has a higher luster and yields smoother, more continuous cleavage surfaces than does pyroxene. Some varieties of amphibole have long, needle-like crystals, resulting in a fibrous structure. Pyroxene does not occur in this form.

It will bear repeating that the presence of well-formed crystal faces or cleavage surfaces is essential in order to distinguish between pyroxene and amphibole in hand specimens.

Occurrence. Amphibole is an abundant rock-making mineral, occurring in both igneous and metamorphic rocks.

Hornblende is a common dark variety of amphibole.

Pyroxene and amphibole together with biotite are the common dark constituents of many rocks. The first two can be distinguished from biotite by the fact that they occur in prismatic crystals that can not be divided into thin elastic flakes; that is, they lack the perfect cleavage of the micas. If present as small grains in a rock, they lack the high luster characteristic of flakes of biotite. They can be distinguished from chlorite by their much greater hardness as well as by their form and lack of micaceous cleavage.

APPENDIX B

ROCKS

> If we would know the life history of our planet, we must learn the origin, structural relations, and composition of our rocks. We must discover the forces—chemical and physical—which work in and upon them, and we must see *how* they work.—George Huntington Williams.

Rocks are the materials composing the units of which the Earth's outer shell is built. It is not possible to give a simple all-inclusive definition of rock. Most rocks are made up of grains of minerals of one or more kinds. Some, however, are made up of fragments of older rocks.

Rocks differ widely in appearance and other properties according to the constituents present, the number of these constituents, their relative abundance, their size, and the pattern formed by their arrangement and association. The kinds of rocks are many, but, if classified according to the ways in which they were formed, they fall into three major

I. Igneous rocks, formed by solidification of molten rock matter, as exemplified by the rocks formed by the cooling of lava poured out from a volcano.
II. Sedimentary rocks, most of which were formed by their substance settling as sediment from a body of water.
III. Metamorphic rocks formed from pre-existing rocks by the development of new characters as the result of heat, deformation, or other geologic agents acting on the pre-existing rocks within the Earth's crust.

Every rock carries within itself more or less evidence of its mode of origin. As one of the prime purposes of geology is to determine the constitution, structure, and history of the Earth's crust, the recognition of rocks and the ability to interpret their meaning is of fundamental

CHARACTERS USED IN IDENTIFYING ROCKS

The properties most useful in identifying rocks are *structure, texture, hardness,* and *fracture.*

Structure is the term used for the larger features of the construction and arrangement of the parts of rocks. Layering is such a feature; it generally indicates sedimentary origin. If a rock contains spherical or almond-shaped cavities called *vesicles* ("blowholes" formed by expansion of gases liberated from the surrounding molten rock matter), it is said to have a *vesicular* structure; such a structure indicates that the rock having it is of igneous origin. If the vesicles become filled with minerals, the resulting structure is called *amygdaloidal* (Fig. 359).

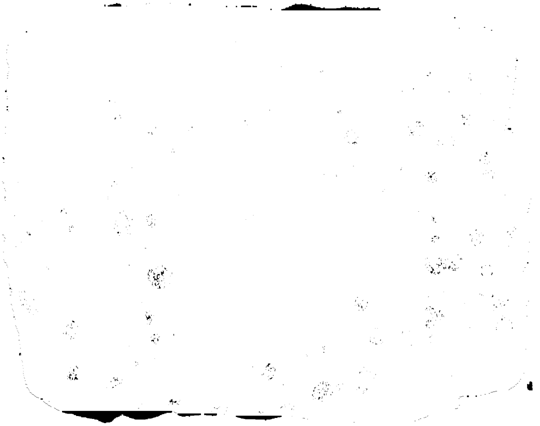


FIG. 359. Amygdaloidal structure. The cavities, originally spherical and almond-shaped, became filled with newly deposited minerals forming *amygdales;* the resulting rock is an *amygdaloid.* The circular cross-sections of the amygdales contrast with the rectilinear boundaries of the phenocrysts in porphyries (Fig. 361).

Texture is the appearance of a rock as determined by the size, shape, and arrangement of its constituent mineral grains. The magnitude of the grains determines the *grain size* of the rock: if the grains are as large as peas, the rock is *coarse grained* in texture; if they are the size of those in granulated sugar, the rock is *fine grained.* If a rock is made up of grains large enough to be recognized by the unaided eye, it is said to be *phanerocrystalline,* or more shortly, *phaneric* (făn-ĕ'rĭk). If the grains are so small that they can not be distinguished as individuals by the unaided eye and the rock seems to be a homogeneous substance, the rock is said to be *aphanitic.*

The shape of the mineral grains and their arrangement with respect to one another produce a characteristic pattern, called the *fabric* of the rock. For example, a rock composed of grains of about one size has an *equigranular* fabric (Fig. 360), and a rock in which the grains differ in size has an *inequigranular* fabric. A common form of inequigranular fabric is called *porphyritic,* in which large conspicuous crystals are inclosed in a fine-grained or aphanitic matrix. The matrix is called the

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FIG. 360. Equigranular igneous rock. Granite from Glacier Point, Yosemite Valley, California.

FIG. 361. Porphyry; contains many well-formed phenocrysts (the white crystals of feldspar), which are inclosed in an aphanitic groundmass.

groundmass, and the large conspicuous crystals inclosed in it are called *phenocrysts* (fē′nō-krists). A strikingly porphyritic rock is shown in Fig. 361. The explanation of the origin of the porphyritic fabric is given on page 296.

Certain textures are of definite help in identifying rocks. The texture of a granite, which is so distinctive that it is termed the *granitic texture,* proves not only that the granite is of igneous origin, but also that it formed under conditions of slow undisturbed cooling. A rock of glassy texture is also of igneous origin, but the glassy texture shows that the rock, unlike granite, was formed by the rapid solidification of molten rock matter, for a natural glass is the result of drastic chilling. *Clastic texture,* which occurs in rocks that are made up of angular or more or less rounded fragments of minerals and rocks, is characteristic of many sedimentary rocks. Other textures are described in connection with particular rocks.

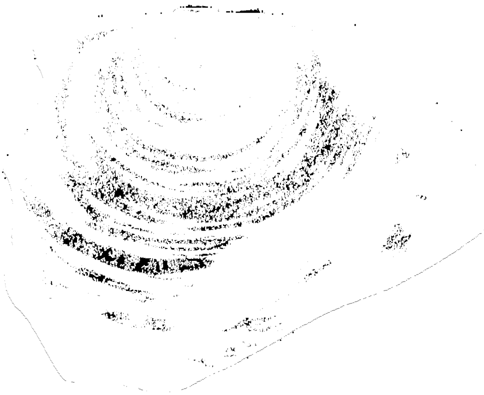

FIG. 362. Conchoidal fracture in volcanic glass.

Hardness is useful in distinguishing between certain kinds of rocks. Many rocks resemble limestone, but the test for hardness with the knife point serves at once to distinguish a limestone, whose hardness is 3, from the much harder rocks that resemble it.

Fracture is a less useful property. However, perfect conchoidal fracture characterizes volcanic glasses (Fig. 362); and a semi-conchoidal fracture yielding shell-like fragments characterizes shales. Most metamorphic rocks tend to split into slabs or thin flakes, and this property is a valuable aid in their recognition.

IGNEOUS ROCKS

Since rocks formed by the congelation of molten rock matter vary in texture and in composition, these two features are used in classifying the igneous rocks. By taking texture as the criterion for subdividing the igneous rocks, we obtain five major classes:

I. *Equigranular phanerites,* in which all the component minerals are about the same in size and are large enough to be identified by the eye alone or aided by a pocket lens.
II. *Porphyritic-granular,* in which certain minerals—the phenocrysts—by virtue of their large size contrast conspicuously with those which surround them, thus forming a porphyry having an equigranular groundmass.
III. *Porphyritic-aphanitic,* in which the large crystals—the phenocrysts—are set in an aphanitic groundmass.
IV. *Aphanites,* in which all the constituents are indistinguishable by the unaided eye.
V. *Glasses,* in which few or none of the constituents have crystallized.

This order from Class I to Class V marks in a general way the decreasing amount of easily recognizable minerals in rocks: in Class I, all the constituents can be easily distinguished as individual grains by the unaided eye; in II, the phenocrysts can be easily distinguished, but the constituents of the groundmass less readily; in III, the phenocrysts alone are distinguishable; and in IV, none of the constituents can be recognized.

Each major class is then subdivided on the basis of its mineral composition—on the kinds of minerals present and the proportions in which they occur. To these subdivisions the actual rock names are given. For example, an equigranular rock composed of feldspar, quartz, and generally also a dark mineral, as a rule biotite, is called *granite.*

By applying these principles, the following classification of igneous rocks is obtained, as shown in the accompanying table.

Remarks on the Table. The glassy rocks, which are rare, being left out of account, the following remarks should aid in making clear the classification of igneous rocks shown in the table.

All rocks in the same horizontal row have the same texture.

All rocks in the same vertical column have essentially the same chemical composition; for example, granite, granite porphyry, and rhyolite

Table of Igneous Rocks

DECREASING GRAIN SIZE ↓	Major Classes (based on texture)	Subdivisions of Major Classes (based on mineral composition) Generally increasing proportion of dark minerals→			
		Light-colored minerals, chiefly feldspar, predominate		Dark minerals predominate	Dark minerals entirely
I	Equigranular, coarse grained (phaneric)	Granite (has quartz)	Diorite (has no quartz)	Gabbro Dolerite (grain size is intermediate between that of gabbro and basalt)	Peridotite Hornblendite Pyroxenite
II	Porphyritic, with generally fine-grained granular groundmass	Granite Porphyry (has quartz)	Diorite Porphyry (has no quartz)	Gabbro Porphyry	
III	Porphyritic, with aphanitic groundmass	Rhyolite (has phenocrysts of quartz)	Andesite (has no phenocrysts of quartz)	Basalt	
IV	Nonporphyritic, aphanitic	Felsite			
V	Glassy	Obsidian Pitchstone Pumice		Basalt Glass	

are alike in chemical composition. In physical appearance, however, they differ notably: a granite differs somewhat from a granite porphyry, and vastly from a rhyolite. Yet it is possible to assemble a series of specimens that will bridge by imperceptible gradations the vast gap between granite and rhyolite. The differences between granite, granite porphyry, and rhyolite are mainly the results of different rates of cooling of magmas of identical composition. When such a magma cools slowly within the Earth's crust it forms granite, whereas the same magma, if erupted on the Earth's surface where it would be drastically chilled, would solidify as rhyolite. In the same way a magma that would yield diorite in depth would, if erupted at the surface, solidify as andesite; and a magma that yields gabbro in depth solidifies as basalt on the Earth's surface. These facts can be epitomized thus: rhyolite is the extrusive equivalent of granite; andesite is the extrusive equivalent of diorite; and basalt is the extrusive equivalent of gabbro.

The rocks in which the light-colored minerals predominate are light in color and light in weight; that is, they are of low specific gravity.

The rocks in which the dark (ferromagnesian) minerals predominate are dark in color and heavy in weight. The range in specific gravity—from 2.67 for the average granite to 3.0 for gabbro—is not large, but is sufficient after experience to serve as an aid to identification.

Although in the table each rock is in a separate compartment, in Nature no rock variety is as sharply delimited from its neighbors as it seems to be in the table. For example, there are transitional varieties between granite and diorite and between granite and granite porphyry. No hard and fast boundaries set off any of the so-called rock species. These facts often make it difficult to classify a given rock. Further difficulties are presented by the finer-grained rocks, especially the aphanitic group. When accurate identification of a rock becomes a matter of high importance, recourse must be had to the microscope.

EQUIGRANULAR PHANERITES

An equigranular phanerite is the result of slow cooling, or of slow cooling combined with retention of the gas content in the cooling magma until a late stage in its solidification. Typically, then, the phanerites occur in deep-seated intrusive bodies, especially in batholiths (p. 289). In such deep-seated masses cooling was necessarily long protracted, and the pressure was sufficient to keep the gases within the magma and to allow them to exercise their power of promoting coarse crystallization. Hence these rocks are often called *deep-seated rocks* or *plutonic rocks* in fanciful reference to the realm in which they originated.

Granite. As may be seen from the scheme of classification, *granite* is composed largely of quartz and feldspar (mainly of the variety orthoclase). It contains also as a rule some mica, generally the black mica biotite; less commonly it contains hornblende, or hornblende and biotite together. All these component minerals are roughly of the same size, and large enough to be recognizable by the unaided eye. Hence granite is said to be equigranular and phaneric. The minerals began to separate from the magma in a definite order: first the dark minerals, hornblende and biotite; then the feldspars; and last the quartz. The dark minerals, being the first to crystallize, were not hampered in their growth by the presence of any neighbors, and so are generally in the form of sharply defined crystals; and the feldspars, having begun to grow later, are less well crystallized, for where they abutted upon the earlier-formed dark minerals their freedom to grow was hampered. As the quartz was the last mineral to separate from the magma, it had to take what space was left, and it is therefore molded around the earlier minerals and occupies the angular interspaces between them. This habit of the quartz produces an intimately interpenetrating and interlocking arrangement. This interlocking equigranular phaneric texture is so characteristic of granites that it is called for short *granitic texture*. It serves to distinguish rocks of Class I from all others.

Closely allied to granite is *granite pegmatite*, which is composed chiefly of potassium feldspar and quartz and has an abnormally coarse grain.

The average granite contains 60 per cent of feldspar, 30 per cent of quartz, and 10 per cent of dark minerals. There are many varieties of granite, based on color, texture, and composition. Its common occurrence is shown in the fact that there are few States in the Union or Provinces in Canada that do not contain exposures of granite; and its use as a building and monumental stone is well known.

Syenite is much like granite in composition but differs in containing little or no quartz. It is not a common rock, nor does it as a rule occur in large masses compared with the enormous bodies of granite.

Diorite. *Diorite* is an equigranular igneous rock composed of feldspar and one or more dark minerals, in which the feldspar is more abundant than the dark minerals. The feldspar is mainly plagioclase, but unless the characteristic striae on the basal cleavage planes of the plagioclase can be seen it is generally difficult to recognize the plagioclase with the unaided eye. The dark minerals are biotite, hornblende, or pyroxene, occurring either singly or together.

Gabbro. The average *gabbro* differs from the average diorite in that the feldspar is subordinate and the dark minerals predominate. Hornblende, pyroxene, and olivine are the common dark minerals, occurring singly or together; biotite, though present in some gabbros, is distinctly uncommon. Because of the prevalence of dark minerals, gabbro is dark and of high specific gravity. *Dolerite* is a convenient term for the basic rocks that in grain size are intermediate between basalts and gabbros.

Peridotite. *Peridotite* (pĕr′ĭ-dō-tīt) is composed wholly of ferromagnesian minerals, with olivine predominating. It is not common and generally occurs as minor intrusive bodies—dikes, sills, and stocks (p. 292). It is interesting and important, however, as being the source of ores of chromium, nickel, and platinum, and of some of the ores of iron. It is generally dark or black, and heavy from the large amount of iron-bearing minerals present.

A notable feature of peridotite is its tendency to alter to a dark-green rock, *serpentinite*. In this change the minerals of the peridotite combine with some 14 per cent of water, and this added water causes a large expansion of volume, which in turn causes much internal movement in the attempt of the mass to accommodate its increased volume. As a result the serpentinite, or "serpentinized peridotite," is traversed by countless smooth shiny surfaces known as *slickensides* (p. 374). Most peridotite masses the world over are more or less serpentinized.

Pyroxenite, as its name implies, is composed wholly of pyroxene, and *hornblendite* consists entirely of hornblende. As a rule these rocks form bodies of small size; nevertheless, in places, as at the remarkable platinum deposits discovered in South Africa, pyroxenite occurs in vast volume.

PORPHYRITIC-GRANULAR ROCKS

The distinguishing feature of rocks of this class is that they contain phenocrysts imbedded in a groundmass so coarse grained that its component minerals can be recognized by the unaided eye. The phenocrysts in most of these porphyries are abundant, making up half or more of the bulk of the rock. If they exceed 75 per cent of the volume of the rock, they are so crowded that the porphyry is indistinguishable from the corresponding granular rock.

Porphyries are common as minor intrusive bodies: as dikes, sills, volcanic necks, stocks, and laccoliths (p. 287). They do not occur as batholiths.

Granite Porphyry, Diorite Porphyry, etc. The typical *granite porphyry* contains conspicuous crystals of feldspar, quartz, and biotite, which are set in a granitic groundmass. As its name implies, its composition is like that of granite; it differs from granite in having phenocrysts, and on the average the grain of the groundmass is finer than the grain of the average granite.

Diorite porphyry differs from granite porphyry in containing no quartz phenocrysts and in having many phenocrysts of plagioclase feldspar.

PORPHYRITIC-APHANITIC ROCKS

Rocks of this class are generally of volcanic origin. Extruded upon the Earth's surface, the magmas from which they were formed have cooled rapidly. These rocks are characterized by the occurrence of porphyritic crystals set in a groundmass that is so fine grained as not to be resolvable by the unaided eye or else is partly or wholly glassy.

Rhyolite. *Rhyolite* represents the aphanitic lava form of the magma that at depth consolidates as granite. It contains phenocrysts of feldspar, quartz, and biotite, and rarely of hornblende. The phenocrysts range in abundance within the widest limits, so that there is a complete transition from nonporphyritic to highly porphyritic rhyolite. If the amount of phenocrysts exceeds 25 per cent of the volume, the rock is by some called a *rhyolite porphyry*. The colors range from white to gray, pink, red, and purple. Rhyolites and andesites with inconspicuous phenocrysts or with few or no phenocrysts are termed *felsites*.

Andesite. *Andesites* are of many colors, but in general they are darker than the rhyolites; dark gray and black are common. They are transitional on the one hand into rhyolites; on the other, into basalts. The average or typical andesite occupies the intermediate position. The darker andesites are of basaltic appearance, but unlike basalts their freshly broken thin edges are translucent when held in bright light. The phenocrysts in andesite commonly consist of striated feldspar and one or more dark minerals (hornblende, pyroxene, or biotite). Quartz phenocrysts are absent (distinction from rhyolite). Andesite containing many prominent phenocrysts is by some called *andesite porphyry*.

Andesite and andesite porphyry are enormously abundant among the extrusive rocks of the globe. They are the chief products of the volcanoes that form the "circle of fire" surrounding the Pacific Ocean. In fact, their prevalence in the *Andes* of South America caused them to be given their name. On account of their great abundance and their many differences in color, texture, and mineral composition the variety of andesitic rocks is legion.

Felsite. The difficulty of discriminating between rhyolite and andesite that are devoid of phenocrysts makes it necessary to use an elastic noncommittal name. For the light-colored rocks of this class, namely those which are white, light to medium gray, light pink to dark red, pale yellow or brown, purple, or light green, rather than dark green, dark gray, dark brown, or black, the term *felsite* is convenient.

Some felsites that appear in hand specimens almost as dark as basalts should be examined on their thin edges, where the rock is seen to be almost white from transmitted light. Basalt specimens are dark even on thin edges.

Basalt. Lavas that are dark gray, dark green, brown, or black are termed *basalt*, the common extrusive equivalent of the basic magmas. Basalts are either compact or vesicular (p. 309). If the vesicles have become filled with some mineral, such as calcite, chlorite, or quartz, the fillings are called *amygdales*, and the rock is termed an *amygdaloidal basalt*. Many basalts are without phenocrysts, but others have abundant conspicuous phenocrysts, consisting of olivine, or pyroxene, or less commonly plagioclase, or some combination of these. Therefore, in the Table of Igneous Rocks basalt is shown to fall in both Classes III and IV. The phenocrysts are hard and have straight, clean-cut boundaries, whereas most amygdales are soft and have irregular, roundish, or elliptical boundaries.

Basalt is by far the most voluminous of the extrusive rocks. The enormous tracts in western America, India, and elsewhere that have been flooded by outpourings of basalt are mentioned in Chapter 14.

Dolerite is the name given to the coarser-grained basalts, in which the grains are large enough so that the constituent minerals are nearly or quite recognizable. There is no hard-and-fast line between basalt and dolerite on the one hand and dolerite and gabbro on the other.

GLASSY ROCKS

Volcanic glass occurs as thin crusts on the surfaces of lava flows and as lava flows that have cooled rapidly. Most glasses are the products of the chilling of silicic magmas (p. 299). Brilliantly lustrous volcanic glass is called *obsidian*, and the duller and more pitchy-lustered variety is *pitchstone*. *Pumice* is frothed glass. Natural glasses, like the obsidian of Obsidian Cliff, Yellowstone National Park, commonly contain crystallized minerals in the form of small spheres which have a radiating or spokelike structure and are known as *spherulites*.

Obsidians are generally dark colored to black. Yet many of them correspond in chemical composition to rhyolite and granite; hence they seem to contradict the rule that nearly all rocks of siliceous composition are light colored. However, if a piece of black obsidian is chipped to a thin edge it transmits the light and loses much of its dark appearance. The deep coloring is the result of the uniform distribution throughout the glass of a relatively small amount of dark material.

Basalt glass is of rare occurrence. Its formation requires extremely drastic chilling and rapid solidification of basaltic magma.

SEDIMENTARY ROCKS

Sedimentary rocks are formed principally in two ways. Some are formed by the accumulation of fragments derived from older rocks: detritus consisting of rock and mineral particles is carried away from its source rocks by water, wind, or ice, is eventually deposited, and becomes hardened into rock. *Detrital rocks* thus formed are classified according to the size of the constituent grains of detritus. The second principal class of sedimentary rocks is made of material formerly dissolved in the sea (and to a lesser extent in lakes), from which it has separated either in the form of the shells of organisms or as chemical precipitates; rocks of this origin are classified according to composition

By far the most abundant sedimentary rocks are shale, sandstone, limestone, and conglomerate.

Conglomerate consists of gravel that has become firmly cemented. The pebbles in it are more or less rounded (Fig. 363), having become water worn by abrasion during stream transport or by buffeting by waves in the shore zone. They consist of rocks of any kind, but generally of some durable material, such as quartz and quartzite. The filling in the interspaces between the pebbles consists of sand and a cement, which may be silica, clay, iron oxide, or calcium carbonate.

FIG. 363. Conglomerate, containing well-rounded pebbles. About one-half natural size.

Breccia (brĕt′shĭa) is like conglomerate, except that most of the fragments, instead of being water-worn pebbles, are angular, with sharp edges and unworn corners. No sharp demarcation separates breccia and conglomerate, since the distinction between them rests on the shape of the constituent fragments, which ranges from complete angularity through subangularity to complete roundness. Breccia, in brief, is a coarse angular detrital rock. A firmly cemented alluvial breccia is termed *fanglomerate* (Fig. 178, p. 254), and that of glacial origin is termed *tillite* (p. 176).

Sandstone. *Color.* A wide range of colors: gray, yellow (buff and tawny), red, and brown are most common, but green and other tints occur.

Physical Properties. Consists of firmly cemented sand grains. Coarse sandstone grades into conglomerate; on the other hand, fine-grained sandstone grades into siltstone, the lower limit of grain size for a sandstone being that at which the individual grains are not distinguishable by the unaided eye. The grains are chiefly particles of quartz or of other durable material. The substance filling more or less completely the interspaces between the grains is the *cement*. The strength and durability of a sandstone depend on the nature of its cement, and the porosity depends on the extent to which the spaces between the grains have not been filled. As in conglomerates, the cement differs in different sandstones. A siliceous cement produces the strongest and most durable sandstone.

Sandstone when fractured breaks around the grains instead of through them, because the grains are stronger than the cement. Therefore sandstone on its broken surfaces has a gritty feel. In quartzites (p. 576), however, the grains are so firmly cemented that fracturing takes place across the grains instead of around them.

Arkose is a feldspar-bearing sandstone more or less resembling a granite. The feldspar can be recognized by its cleavage planes, which reflect flashes of light as the specimen is turned from side to side while being examined. If an arkose contains biotite, its resemblance to granite is greatly heightened. Arkose and granite can be distinguished by the fact that in the arkose the quartz is in angular or subangular particles, instead of being molded around the feldspar as it is in granite.

Siltstone. Siltstone is silt that has been converted into rock. It is intermediate in grain size between sandstone and shale.

Shale. *Color.* Gray in various shades perhaps most common; but red and pink in many shades, brown, buff, green, and black are also common.

Physical Properties. Shale is so fine grained as to appear homogeneous to the unaided eye. It is soft enough to be easily scratched. Typically it has a smooth and almost greasy feel; but a small amount of fine sand in its composition makes the shale somewhat gritty. A semiconchoidal fracture is common in shale, causing it to split parallel to the bedding into thin shell-like plates or fragments.

Shale is essentially clay that has been converted into rock. It is therefore high in clayey constituents and is termed an *argillaceous* rock (Latin *argilla*, clay).